PETROLOGY AND BURIAL DIAGENESIS OF PLIO-PLEISTOCENE SEDIMENTS, NORTHERN GULF OF MEXICO

APPROVED BY THE SUPERVISORY COMMITTEE:

Wm. L. Fisher

W. E. Galloway

Earle F. McBride

Edward D. Pittman

PETROLOGY AND BURIAL DIAGENESIS OF PLIO-PLEISTOCENE SEDIMENTS, NORTHERN GULF OF MEXICO

by

Kitty Lou Milliken, B.A., M.A.

DISSERTATION

Presented to the Faculty of the Graduate School of

The University of Texas at Austin

in Partial Fullfillment

of the Requirements

for the Degree of

DOCTOR OF PHILOSOPHY

THE UNIVERISTY OF TEXAS AT AUSTIN

MAY, 1985

ACKNOWLEDGMENTS

I would like to acknowledge Dr. Lynton S. Land, best-of-supervisors, for the time and enthusiasm he contributed to this project and others during my stay at U.T. This dissertation is very much a product of his 'holistic' view of basinal diagenesis (Of course, I claim entirely for myself any weaknesses in this interpretation of that view). One key ingredient in my rather long stay at U.T. has been the entertaining and enlightening presence of Dr. Robert L. Folk, who, among other things, taught me to read. Dr. Earl F. McBride also worked to further my education by offering alternative, and sometimes opposing, ideas of what sedimentary rocks are all about. For almost a decade, these three gentlemen have provided me with an academic committee without compare and I thank them.

Also serving on the final dissertation committee, Drs. W. E. Galloway, Bureau of Economic Geology and E. D. Pittman, Amoco Production Company, generously shared their knowledge of diagenesis and their editorial skills to the great benefit of the final draft. Another superb editorial job was done by Wendy MacPherson, student editor.

Samples for this project were provided by Amoco,

Marathon, Mobil, Shell, and Texaco. I would like to thank the people at these companies not only for samples and well data but also the work space and hospitality they provided at their core storage facilities. In particular, I would like to acknowledge several people who made special efforts to facilitate my sampling of Plio-Pleistocene sediments: Selena Dixon of Mobil; Dennis Prezbindowski, Dick Larese, and Ed Pittman of Amoco; Michele Bishop of Marathon; John Branch of Shell; and Richard George of Texaco. Time on the electron microprobe at the Cities Service research lab was kindly arranged for me by Ken Helmold.

Financial support for this dissertation was provided through a teaching assistantship in the Department of Geological Sciences, several small grants for travel from the Geology Foundation at the University of Texas, a research grant from the Gulf Coast Association of Geological Societies, a one-year fellowship sponsored by Tenneco, and two years of research support from Texaco. I am grateful for the research opportunities made possible by these various forms of support.

Acknowledgments would not be complete without

mentioning some of my numerous fellow-students and friends of diagenesis. Shirley Dutton, Mike Boyles, Steve Fisher, Jim Anderson, Dave Budd, Steve Johansen, Ralph Kugler, Tim Jackson, Don Miser, Paul Lundegard, Larry Mack (provider of the strontium isotopic analyses), Ellen Graber, Paul Gold, Wendy MacPherson, Randy Farr, and Paul Blanchard (listed very roughly in the order I got to know them) have all been wonderful friends who've taught me much. Special friend Steve Seni taught me the most and kept me going through it all.

PETROLOGY AND BURIAL DIAGENESIS OF PLIO-PLEISTOCENE SEDIMENTS, NORTHERN GULF OF MEXICO

Publication No. _____

Kitty Lou Milliken, Ph.D.
The University of Texas at Austin, 1985

Supervising Professor: Lynton S. Land

Plio-Pleistocene sediments and rocks beneath the Louisiana and adjacent Texas shelves are the youngest of several thick packages of terrigenous sediment which prograded into the Gulf of Mexico during the Cenozoic. Comparison of diagenesis in these young sediments (more than 300 samples from 45 wells on the Louisiana-Texas shelf) to diagenesis of older Cenozoic rocks at similar burial depths elsewhere along the Gulf margin confirms that diagenesis is not strictly analogous among the various Cenozoic units. There has been an evolution of diagenesis during filling of the Gulf of Mexico. Differences in diagenesis cannot be attributed to differences in bulk mineralogy of the sands because Plio-Pleistocene sands are lithic arkoses and feldspathic litharenites with essentially the same QFR

proportions as observed in subsurface Eocene and Oligocene sandstones along the Texas coast. Unaltered plagioclase is slightly more calcic (average An 24) than unaltered plagioclase in the older rocks.

Burial diagenesis in Plio-Pleistocene sediments has involved essentially the same processes as observed in the older rocks, but overall, diagenesis has advanced to a lesser degree at any given depth. Cementation by quartz and carbonate, dissolution of potassium feldspar and heavy minerals, albitization of plagioclase, and the transformation of smectite to illite have occurred in Plio-Pleistocene sediments, but cements and altered grains are not volumetrically significant shallower than 4 to 4.5 km. The temperature at which reaction of detrital constituents begins (approximately 90^{0} C) is similar to that observed elsewhere in the Gulf, but the zone of reaction is spread over a greater depth range. The similar temperatures observed for the advent of detrital reactions across the Gulf basin suggest that these processes are more highly dependent upon temperature than upon time and that differences observed among the various units may be attributed, at least in part, to variations in the geothermal gradient.

The degree of detrital grain alteration observed

in these young sediments shows that significant loss of provenance information occurs quite early in the burial history. Alteration in the deep subsurface is very effective in modifying the primary detrital assemblage.

TABLE OF CONTENTS

page

LIST OF FIGURES

LIST OF TABLES

INTRODUCTION

Pliocene and Pleistocene sediments beneath the Louisiana and adjacent Texas shelves are the youngest of the thick wedges of terrigenous debris which have infilled the margins of the Gulf of Mexico since Cretaceous time. Study of petrography and diagenesis of these youngest sands and muds was undertaken to accomplish three related aims.

The first goal was simply descriptive. Information on detrital and authigenic minerals in terrigenous units of Late Cretaceous through Miocene age has been (or is being) compiled. Similar documentation for Plio- Pleistocene sediments is needed to complete this data base. Plio-Pleistocene sediments are of particular interest because they represent material shed from central North America during the variable climatic and tectonic conditions of the latest Tertiary and early Quaternary.

Using the descriptive data, the second goal can be addressed--that is, to understand the diagenetic evolution that has accompanied progressive filling of a continental margin basin. Work accomplished to date shows that there are significant variations in the nature and timing of diagenesis among the different units in the Gulf of Mexico basin (Land, 1984a; Land and

Fisher, in prep.). Plio-Pleistocene deposits represent a temporal end member in this basin-filling sequence and the nature of diagenesis in these sediments will help us understand the causes and controls of diagenetic variation in the basin.

The Gulf of Mexico is perhaps our most thoroughly sampled and studied example of a common basin type. Models for diagenesis in the Gulf may find application in other Mesozoic-Tertiary, salt-intruded, continental margin basins around the world. The simple burial history and the relatively recent, even on-going, diagenesis in Gulf coast sediments enables us to constrain diagenesis in ways that are not possible for much older rocks now exposed as a result of complex tectonic histories. Thus, the diagenetic evolution of the Gulf has implications for modelling diagenesis in many ancient examples of thick terrigenous sediments.

The third goal concerns better documentation of the fundamental processes by which sediments are lithified. Alteration of Plio-Pleistocene sand and mud has occurred at depths ranging from the sediment-water interface to approximately 4.5 km, corresponding to temperatures from about 20^{0} to 120^{0} C. These conditions encompass a realm of diagenesis intermediate between very early diagenesis as described by Berner (1980),

among others, and the longer-term, higher-temperature diagenesis or metamorphism recognized in the hotter parts of deep basins and in geothermal areas (e.g. Elders and others, 1979; Milliken and others, 1981). Processes which cause lithification, including compaction, cementation, and grain alteration, all operate in all diagenetic zones. But, it is in the intermediate realm of diagenesis in particular that materials are in large part transformed from sediments to rocks.

Previous Work on Petrology and Diagenesis of Gulf Coast Units

Compiled in Table 1 is a list of studies on petrology and diagenesis of Gulf Coast terrigenous sediments and rocks in the subsurface of the Texas Gulf margin. Also important for understanding the overall picture of Gulf Coast terrigenous diagenesis are studies on:

1. Underlying Mesozoic carbonate rocks---Loucks (1976), Prezbindowski (1981), Woronick and Land (in press), Moore and Druckman (1981).

2. Pore fluids and their movement--Kharaka and others (1978), Land and Prezbindowski (1981), Morton and others (1981), Lundegard and others (1984).

TABLE 1

Petrologic and diagenetic studies of subsurface terrigenous sediments along the margin of the Gulf of Mexico basin.

Author, date	age of unit(s)	location
Weiss, 1979	Late Cretaceous	south Texas
Slator, 1980	Late Cretaceous	N. Mexico
Merritt, 1980	Late Cretaceous	N. Mexico
Boles and Franks, 1979	Eocene	south Texas
Fisher, 1982	Eocene	central Tx.coast
Land and Fisher (in prep)	Eocene	Texas coast
Loucks and others, 1984	Eocene, Olig.	Texas coast
Boles, 1982	Eocene, Olig.	south Texas
Lindquist, 1977	Oligocene	south Texas
Milliken et al. 1981	Oligocene	central Tx. coast
Land and Milliken, 1981	Oligocene	central Tx. coast
Land, 1984	Oligocene	Texas coast
Hower and others, 1976	Oligocene, Mio.	central Tx. coast
Freed, 1981	Oligocene, Mio.	south Texas
Freed, 1982	Oligocene, Mio.	central Tx. coast
Gold, 1984	Miocene	south Louisiana

3. Temperature and pressure regimes--Sharp and Domenico (1976); Loucks and others (1979); Jones (1969, 1975).

Detailed recitation of lithologic features and processes described in these papers is deferred until later sections where this information can be more logically presented together with similar data on the Plio-Pleistocene.

However, the following conclusions, generalized from the above-cited work, should be mentioned.

1. Diagenesis involves a combination of compaction, cementation, and grain alteration. At relatively shallow depths (i.e. less than 1500 m to less than 3000 m, depending on the formation), cementation dominates, whereas deeper, grain alteration is, volumetrically, the more important process.

2. In deeper zones, rock/water interaction has been extensive. In many rocks, only detrital quartz has yet to react with pore fluids. Reactions of various mineral phases have markedly affected the bulk composition (isotopic and ionic) of both rocks and waters. The degree of alteration implies that both water/rock ratios and distances of transport have been large (e.g. Land, 1984a).

3. Diagenesis has progressed to a greater degree

and is prominent at shallower depths in older rocks. At the extremes, extensive lithification is found in Cretaceous sandstones at burial depths less than one kilometer, but Plio-Pleistocene sediments are mostly unconsolidated shallower than four kilometers.

Conditions of Diagenesis

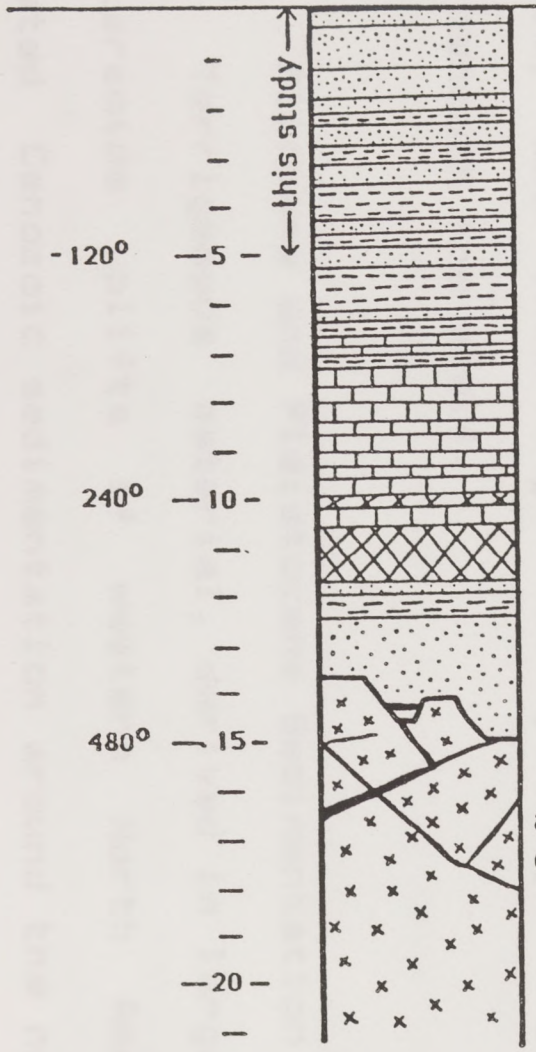
The following is a survey of facts concerning the range of physical and chemical conditions which have affected Plio-Pleistocene sediments in offshore Louisiana. The aim of this section is to set forth, at least to the best of present understanding, the constraints, which must bound any model of Plio-Pleistocene diagenesis. It is convenient to separate this discussion into four parts: temperature, pressure, fluid composition, and depositional history. The first three encompass relatively straight-forward descriptions of conditions observed today in Plio-Pleistocene sediments. Depositional history adds a temporal dimension to our understanding of temperature, pressure, and fluid composition as well as providing some insight into volumes and rates of sediment/water interaction within the basin as a whole.

Depositional history

Pre-Pliocene rocks

Information on Pre-Pliocene rocks beneath the Louisiana shelf is indirect because these rocks are covered with a tremendous thickness of Neogene and Quaternary sediment (on the order of 6 km). Ideas about the nature of these rocks come primarily from extrapolation of information known from drilling and from seismic investigations in the deep Gulf. The relative thicknesses and compositions of post- and pre-Pliocene rocks in the Louisiana shelf region are compared schematically in Figure 1; deep basinal seisimic units are emphasized on this figure because these sediments provide our best approximation for the thickness and composition of sediments onto which the Plio-Pleistocene units were deposited.

The most widely accepted idea is that the material beneath the sedimentary section is thinned continental crust which was attenuated during rifting in the Early Mesozoic (e.g. Ladd and others, 1976; Buffler and others, 1981). Support for this interpretation includes plate reconstruction which places North and South America very close prior to early Mesozoic rifting



temp.	maximum thickness (km)	lithology beneath shelf	seismic units of abyssal gulf		Gulf margin equivalents	age
			SIGSBEE	terrigenous turbidites		QUATERNARY
			CINCO DE MAYO	fine-grained terrigenous turbidites		LATE TERTIARY
-120°	5		MEXICAN RIDGES	pelagic carbonates mixed with fine-grained terrigenous turbidites	FRIO	EARLY TERTIARY
					WILCOX	
			CAMPECHE	deep-water pelagic carbonates	EDWARDS	CRETACEOUS
240°	10		CHALLENGER	salt and other evaporites	SMACKOVER LOUANN	JURASSIC
			VIEJO	continental red-beds and evaporites	EAGLE MILLS	TRIASSIC
480°	15		attenuated continental crust	high-rank quartzo-feldspathic metamorphic rocks intruded by basalt		LATE PALEOZOIC and EARLY MESOZOIC
	20					

Figure 1. Schematic profile of lithologies and temperatures beneath the Louisiana shelf. Summarized from Buffler and others, 1981; Schlager and others, in press; and Woodbury and others, 1973.

(e.g. Dickinson and Coney, 1980), seismic interpretation which indicates approximately 12 km of material with crustal velocities in the area of the Mississippi fan (Ibrahim and others,1981), reasonably symmetric distribution of salt and thicker crust on either side of the thin (5 to 6 km) oceanic crust in the central gulf (Buffler and others, 1981), and core samples of early Paleozoic sialic material intruded by Mesozoic basalts in the eastern Gulf (Schlager and others, in press).

The sedimentary section above oceanic basement in the abyssal Gulf has been divided into six seismic stratigraphic units (Ladd and others, 1976);(Figure 1). These units today are truncated at the base of the modern slope by the Sigsbee Scarp, a salt deformation feature of Neogene or younger age (Buffler and others, 1978). Prior to this deformation, however, it appears that the units of the deep Gulf were continuous farther to the north (Worzel and Burk, 1978) and presumably underlie, and interfinger with, Neogene and younger units beneath the shelf.

Pliocene and Pleistocene Sedimentation

Terrigenous material, derived in large part from the Laramide uplifts of western North America, has dominated Cenozoic sedimentation around the northern and

western margins of the Gulf of Mexico (Murray, 1960). The locus of major deposition of this sediment has shifted on this margin several times during the Tertiary, moving from the west to the north beginning in the Miocene (summarized by Martin, 1978; Winker, 1981). Cenozoic deposition in the northern Gulf has been exceedingly rapid and the combined Pliocene and Pleistocene section in some places on the shelf exceeds 6 km (Woodbury and others, 1973). This corresponds to burial rates on the order of 1500 m/m.y. Depositional characteristics of these young sediments are broadly similar to those of older Cenozoic sediments which rim the western Gulf margin. Sediments can be generally divided into two major depositional types: a thick, mud-dominated prodelta and a continental slope sequence that is overlain by a sequence of interbedded sands and shales deposited in more proximal deltaic and fluvial environments (e.g. Norwood and Holland, 1974; Caughey, 1975; c.f. Galloway, 1984). The history of Pleistocene low stands was such that fluvial and estuarine sedimentation several times extended onto the shelf as far as the present 90 m isobath (Shepard, 1960; Parker, 1960).

Structural development in the Louisiana shelf region has been very much dominated by syndepositional

features such as growth faults (e.g. Braunstein and others, 1973; Holland and others, 1980) and salt diapirs which underwent most of their growth during the Late Cenozoic (Spindler, 1977; Johnson, 1980, Stude, 1978).

Offshore, the subsurface Pliocene and Pleistocene sediments are not assigned to formally recognized formations, but rather are divided on the basis of foraminiferal zones. Onshore equivalents of these sediments are however divided into several formations. Pleistocene formations are correlated reasonably well with the glacial stages. Figure 2 summarizes stratigraphic nomenclature for the post-Miocene section of the northern Gulf of Mexico.

Temperature

Geothermal gradients for the Louisiana shelf are available from the A.A.P.G. Geothermal Gradient Map of North America (1976, Region 7). This map indicates that, within the area encompassed by this study, gradients vary mostly from 16^{o} C/km to 20^{o} C/km, reaching as much as 23^{o} C/km in the western part.

In addition, bottom hole temperatures for wells used in this study show linear variation with depth (^{o}C=0.0204 * depth (m) + 23.2 ($\pm 7^{o}$ C); temperatures uncorrected), and plot along gradients that vary only

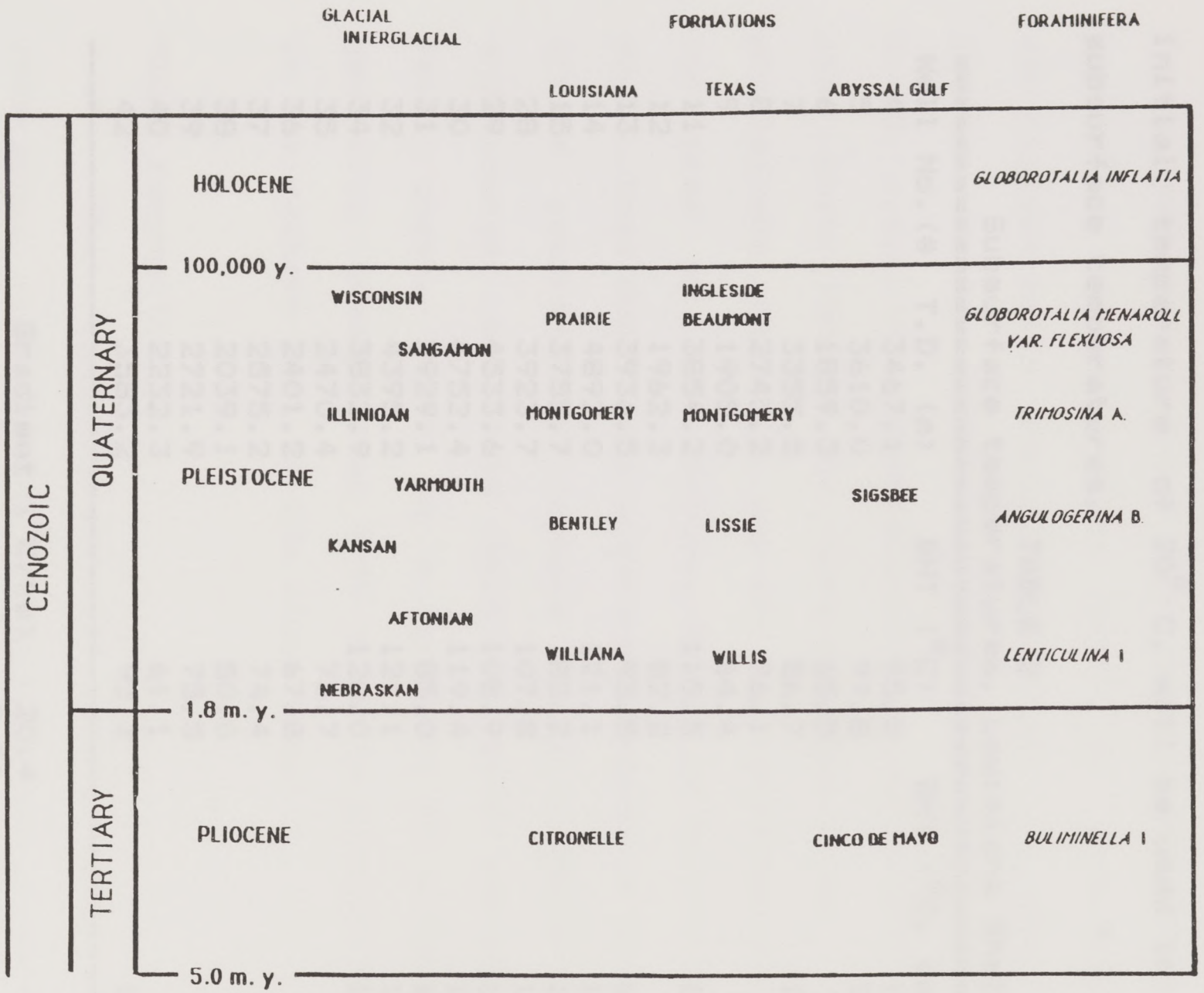


Figure 2. Stratigraphic nomenclature of post-Miocene units of the northern Gulf of Mexico. Summarized from Bernard and LeBlanc, 1965; Beard and Lamb, 1968; Chenoweth and McBride, 1984; Ladd and others, 1976; Waguespack, 1983; and suggestions of W. E. Galloway, personal communication, 1984.

about 1^{o} C/km depending on whether the data are corrected with the equation of Kehle (1971). For simplicity, in this study a gradient of 20^{o} C/km and an initial temperature of 20^{o} C, will be used to calculate subsurface temperatures.

TABLE 2
Subsurface temperatures, Louisiana Shelf

Well No.(#)	T.D. (m)	BHT (^{o}C)	BHT (^{o}C, corrected)*
4	3467.1	95.5	113.9
5	3610.0	97.8	116.1
6	1859.3	65.5	78.3
7	3355.5	86.7	105.0
8	2743.2	76.1	92.8
9	1905.0	64.4	77.2
11	3854.2	115.5	133.9
12	1962.3	82.2	95.5
13	3936.5	95.5	113.9
14	4892.0	121.1	136.1
15	3755.7	82.2	100.6
28	3923.7	107.8	126.1
29	4533.6	108.9	125.5
30	4752.4	119.4	135.6
31	2929.1	85.0	102.2
32	4395.2	121.1	136.1
34	3832.9	120.0	138.3
35	2470.4	76.7	92.2
36	2401.2	67.8	83.3
37	2575.2	74.4	90.5
38	2039.1	50.0	63.9
39	2721.9	75.5	92.2
40	2232.3	61.1	79.9
42	3583.2	93.3	111.7
Gradient (^{o}C/km)		20.4	21.4
r		0.91	0.92

Well locations shown on Figure 2.
* Corrected using Kehle's (1971) empirical equation

Pressure

Little information on subsurface pressure for the offshore Plio-Pleistocene is in the public domain. Top-to-geopressure maps for the Texas on-shore region generally indicate that depth to geopressure rises stratigraphically toward the coast (Loucks and others, 1979). Jones (1969) reports gradients greater than 0.465 p.s.i./ft (geopressure by definition) at depths less than 1500 m in some offshore areas. However, a preliminary isobaric map for the Louisiana shelf region (prepared by D. G. Bebout, Bureau of Economic Geology), shows that depth to a pressure gradient of 0.7 p.s.i./ft ("hard geopressure") occurs generally deeper in Neogene and younger sediments than in older sediments of the Texas coast. Depth to a gradient of 0.7 p.s.i./ft is generally greater than 3000 m and is greater than 4500 m in the eastern part of the study area.

Fluid compositions

The compositions of waters, which may have interacted with sediments beneath the Louisiana shelf, are constrained by what is known about the depositional and burial history. Pore water at the time of deposition of the major, thick deltaic sands most likely was normal seawater. Some mixed meteoric/marine waters

are also possible for this depositional setting. Pore waters in the deeper portions of the older Tertiary units of the Texas coast tend to be isotopically heavier than sea water, a result of interactions between waters and reacting clay minerals (up to $\delta^{18}O$=+7 o/oo SMOW) or carbonate rocks (up to +10 o/oo or more) (e.g. Land and Prezbindowski, 1981). As documented later in this study, clay minerals in Plio-Pleistocene rocks have not reacted to the same extent as observed in these older rocks, so isotopically "heavy" waters might not be expected. Indeed, formation waters sampled in Pleistocene wells on the Louisiana shelf have $\delta^{18}O$ values between +1 and -1 o/oo, the range expected for essentially unmodified depositional fluids, though these waters are highly saline as a result of halite dissolution (Lundegard, 1985).

CHARACTERISITICS OF PLIO-PLEISTOCENE SEDIMENTS

Sampling and methods

Samples for this study were obtained from 45 wells in 30 lease blocks on the Louisiana shelf (Fig. 3). Petrographic data are compiled in Appendix 1. Analytical data for feldspars and clays are given in Appendix 2. The majority of samples collected were cuttings. Cuttings with drilling mud still included were washed in distilled water on a wire screen. This undoubtedly biases these samples toward the more lithified materials from a particular interval, but this was unavoidable.

Point counts of sandstone were made on standard petrographic thin sections treated with Na-cobaltinitrite to stain potassium feldspars and K-rhodizonate to stain calcium plagioclase (staining technique of Houghton, 1980). Two-hundred points were accumulated on grains, resulting in totals between 300 and 500 points for most samples.

Electron microprobe analysis of feldspar was done on grain mounts of the >62 um fraction, obtained by gravity settling. Grain mounts were traversed in a spiral and counts were collected briefly on each grain intersecting the electron beam. Grains which gave count

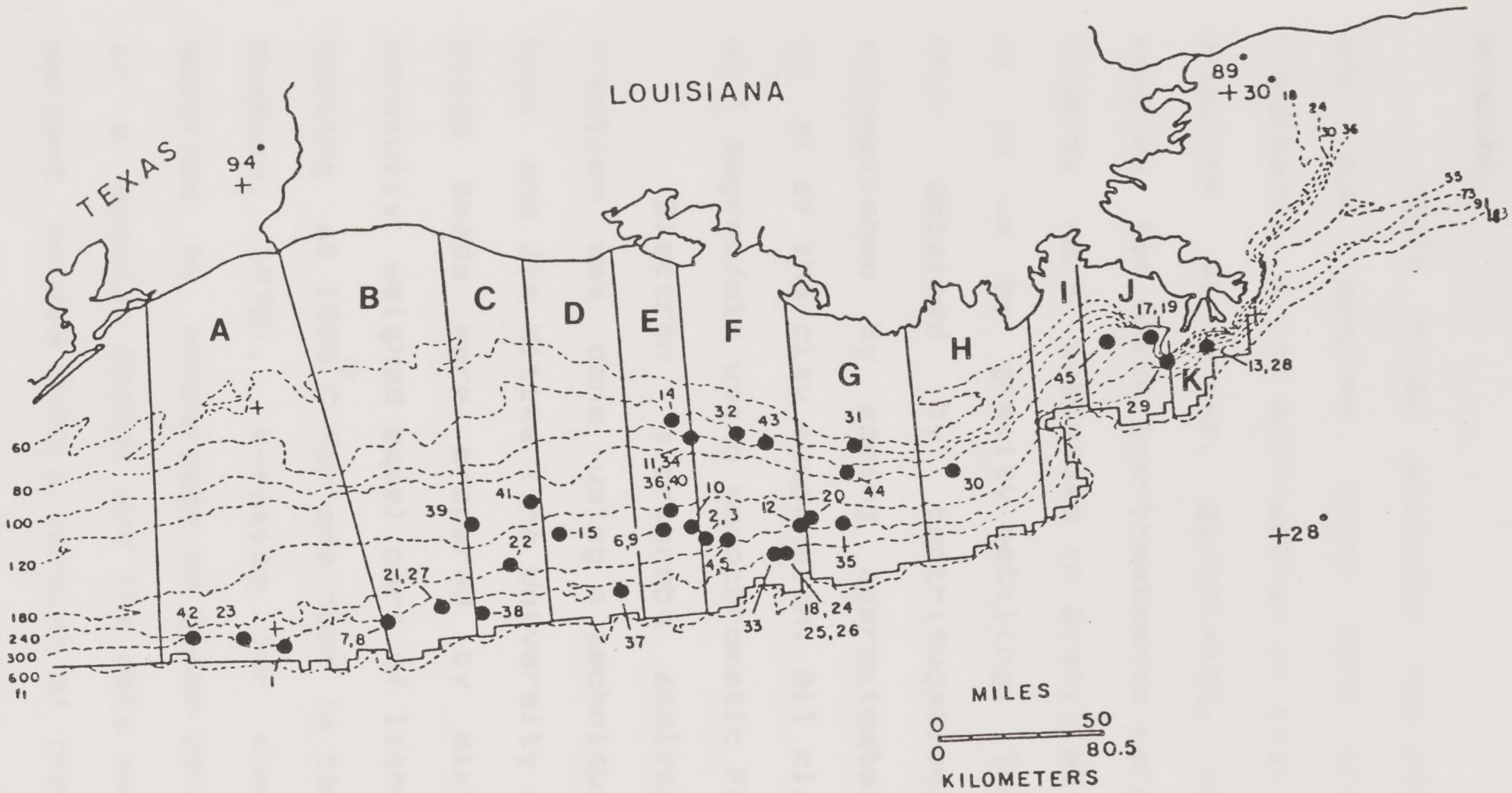


Figure 3. Map of the Louisiana shelf, showing locations of wells used in this study and the modern bathymetry. Well numbers are the same as those listed in Appendix . Letters label federal lease block areas: A. High Island, B. West Cameron, C. East Cameron, D. Vermilion, E. South Marsh Island, F. Eugene Island, G. Ship Shoal, H. South Timbalier, I. Grand Isle, J. West Delta, K. South Pass.

rates reasonable for feldspars were analysed for 30 seconds.

Bulk X-ray analysis was performed on both sand and mud samples which were ground whole without treatment. X-ray diffraction of clay samples was done on oriented air-dried, glycolated, and heated (550^{o} C) samples for which pre-treatments were kept to a minimum. Samples were suspended in distilled water and separated at 10 um by gravity settling. The <1 um fraction was then obtained by centrifugation. Glycolation was accomplished by adding approximately 2 cc of glycol to 40 ml of the clay suspension. All clay patterns were run at 1 degree/min using monochromatic Fe$K\alpha$ radiation.

Electron microprobe analysis of <1 um clay fraction was done using a technique developed by Larry Mack and Jim Wittke (The University of Texas at Austin). Glass beads were prepared by mixing the clay with a carefully weighed equal part of lithium borate flux, and heating at 1000^{o} C for one hour (a technique described by Shapiro, 1975). K-ratios for these beads were then obtained by comparison to glass bead standards prepared in a similar manner and the data were reduced to weight percent oxides with a version of the Bence-Albee program modified to account for weight of the flux (Mack and

Wittke, in prep.).

Stable isotopic analyses of carbonates were done following procedures originally outlined by McCrea (1950). Calcite and dolomite were reacted at 25° C, siderite at 50° C. All samples were routinely X-rayed prior to analysis. Oxygen isotopic analysis of clay samples was done on an aliquot of the <1 um fraction used for clay mineral identification using BrF_5 as an oxidant (Clayton and Mayeda, 1963).

Fractionation factors for the interpretation of stable isotopic data were obtained from the following sources: O'Neil and others (1969) (calcite); Wenner and Taylor (1971) (chlorite); Land (1983) (dolomite); Eslinger and Savin (1973) (illite); Eslinger (1971) (kaolinite); Yeh and Savin (1977) (smectite).

Composition of sands and sandstones

Primary detrital components and their alteration

Bulk compositions

Plio-Pleistocene sands and sandstones of the Louisiana shelf region are primarily lithic arkoses and feldspathic litharenites (classification of Folk, 1980) (Figure 4). The average for all samples counted is 64.9

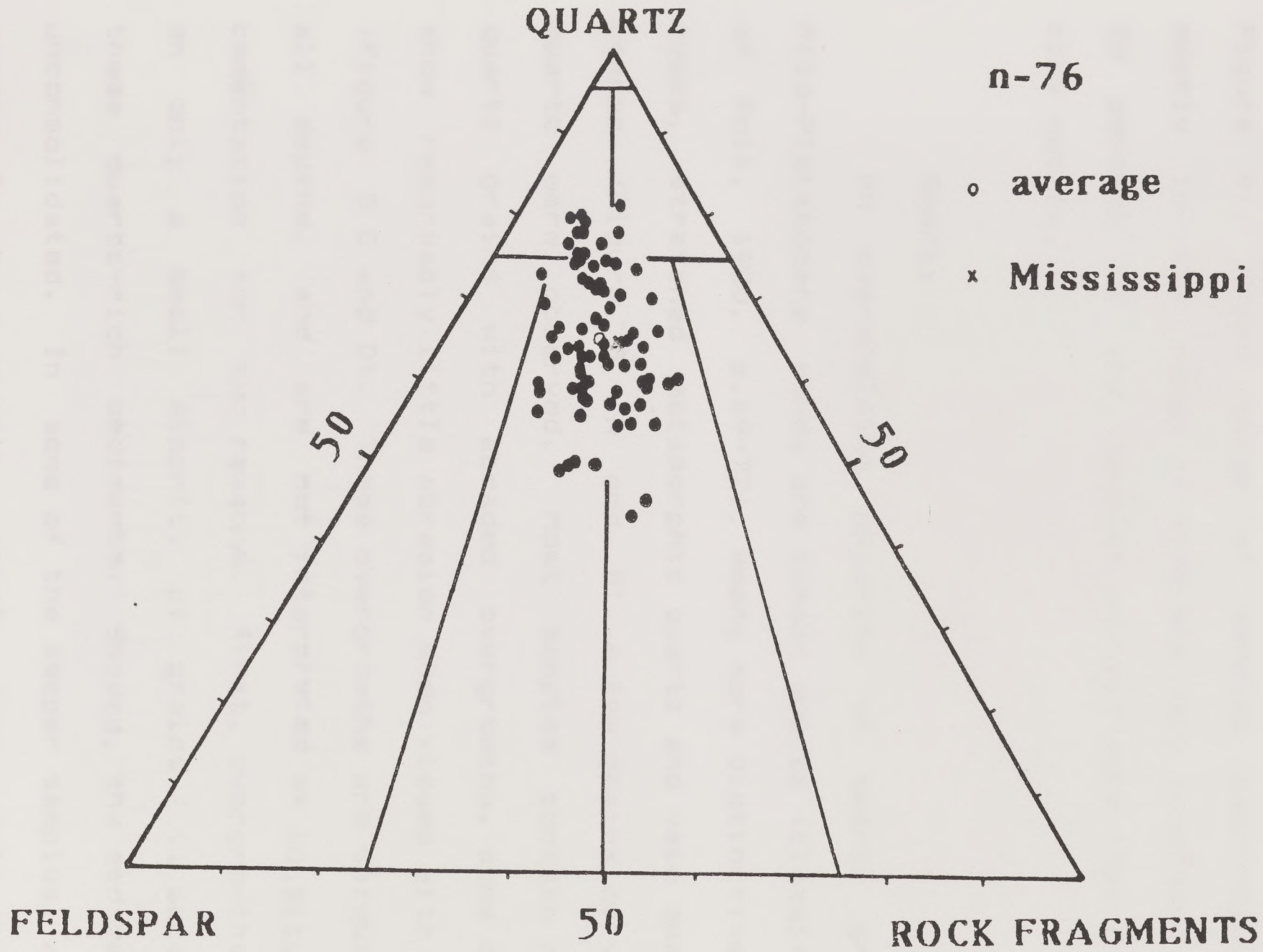


Figure 4. Quartz-feldspar-rock fragment proportions in Plio-Pleistocene sands. Based on the classification of Folk (1980).

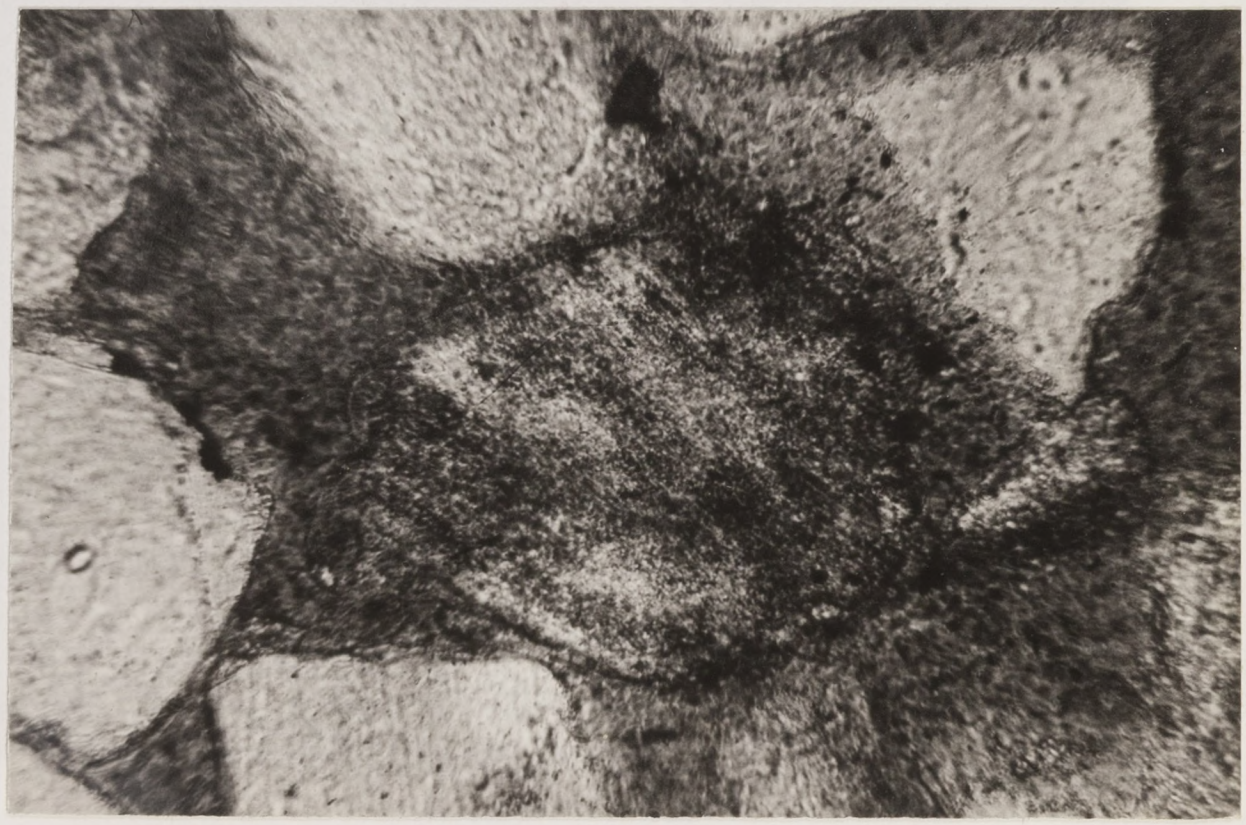
(Q), 18.6 (F), and 16.5 (R). This average is reasonably similar to the composition of a sample of Mississippi River sand, from the area of South Pass (included on Figure 4). Grain sizes of samples point-counted are mostly in the range of fine and very fine sand. About 50 percent of the samples contain more than 5 percent clay matrix. detrital?

Quartz

An overwhelming majority of quartz grains in Plio-Pleistocene sands are common quartz (classification of Folk, 1980, p.69-72). Among more distinctive quartz types, stretched metamorphic quartz and vein quartz are common (Figure 5, A and B). A few grains of volcanic quartz were observed. Most samples contain numerous quartz grains with abraded overgrowths, some of which show remarkably little abrasion when viewed with the SEM (Figure 5 C and D). These overgrowths are ubiquitous at all depths, and are not interpreted as in_situ quartz cementation for two reasons. First, overgrowths appear on only a small minority of grains (<1 percent) in these quartz-rich sediments. Second, the sediments are unconsolidated. In some of the deeper samples, which do appear to have quartz cement, even small amounts of pore-filling material are accompanied by a markedly

A

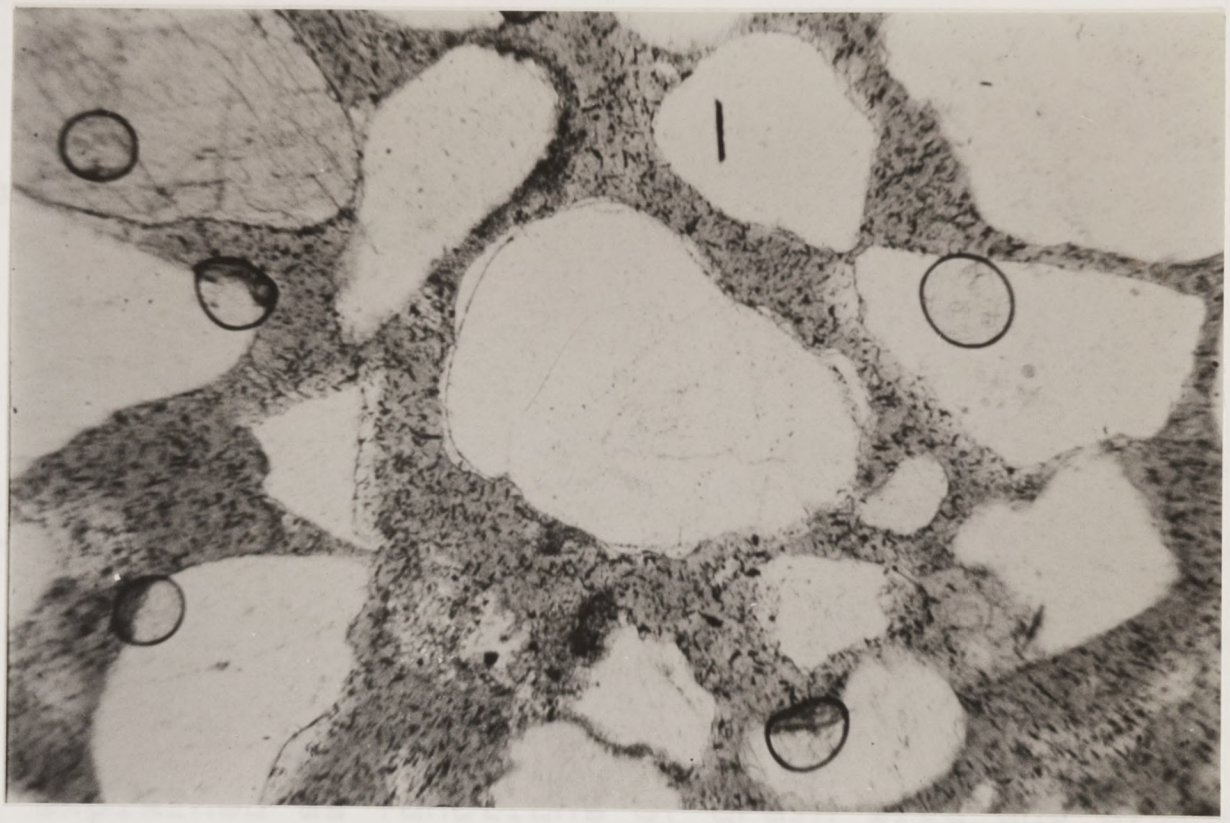
0.1 mm



B

0.05 mm

Figure 5. Quartz types observed in Plio-Pleistocene sands. A. Stretched metamorphic quartz (PP-93, depth=1006 m). Crossed polars. B. Vein quartz. Plane polarized light.



C

D

Figure 5 continued. C. Quartz grain with abraded quartz overgrowth (PP-93, depth=1006 m). Plane polarized light. D. Transported quartz overgrowth observed with the SEM. Abrasion is not apparent (PP-49, depth < 2622 m).

greater degree of lithification.

No evidence of quartz dissolution or replacement was observed; neither have such features been reported for other subsurface units of the Gulf Coast.

Feldspars

Primary detrital feldspar assemblages in Plio-Pleistocene sands reflect compositions expected for Mississippi River detritus (Figure 6). Variations in this assemblage, between samples and with depth, are best explained as functions of grain size and diagenetic modification, as described below.

Potassium feldspar

In thin section, potassium feldspars are composed of approximately equal amounts of microcline and orthoclase, identified on the basis of twinning or lack thereof. These show less evidence of alteration than plagioclase; most grains appear quite fresh or only slightly vacuolized. Sericitization is rare. Despite this overall tendency to appear less altered than plagioclase at all depths, below about 3600 m potassium feldspar is lost by dissolution from the sands (Figure 7). In general, at depths shallower than 3600 m (~90° C), potassium feldspar appears to be quite stable, and

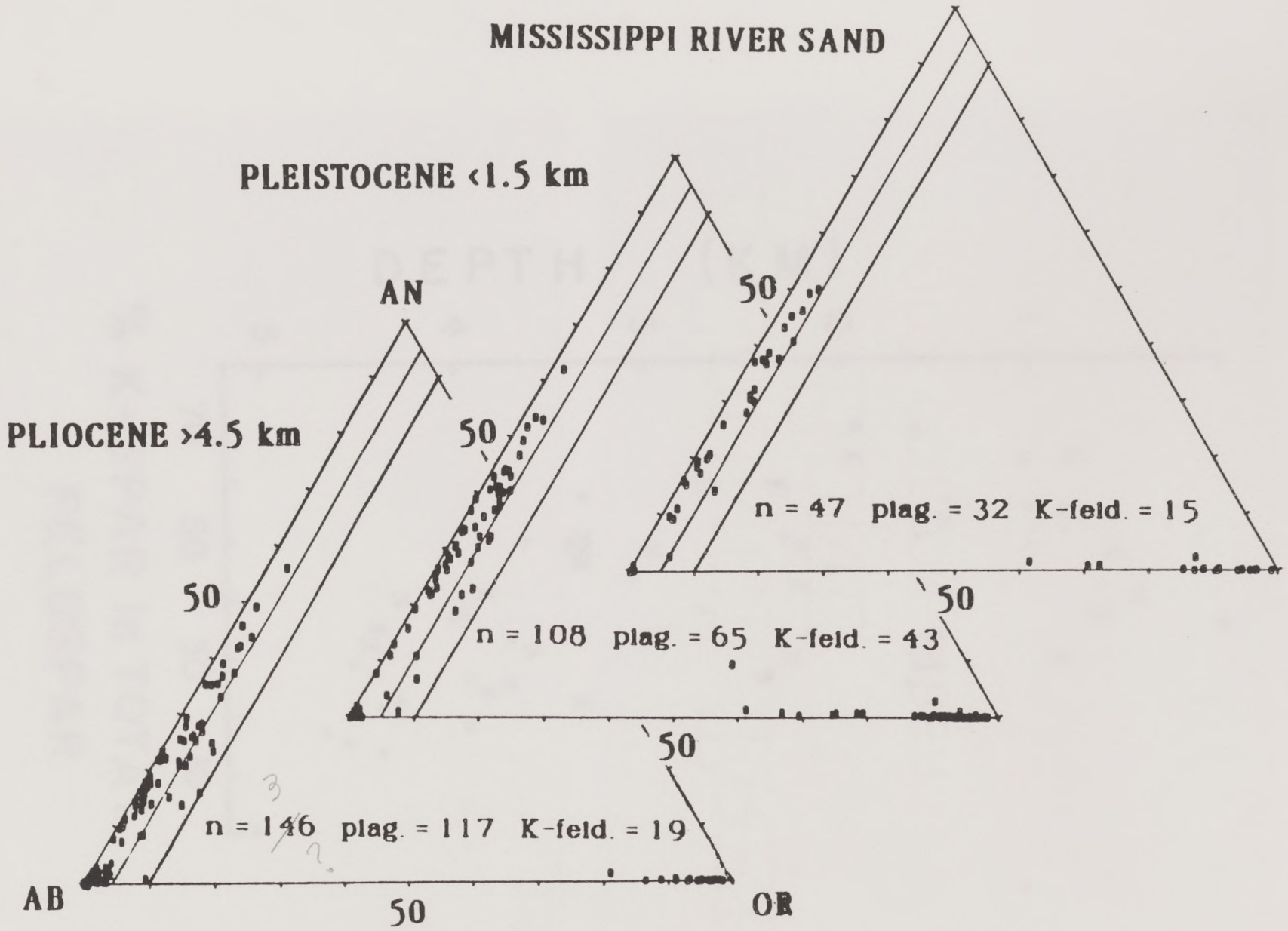


Figure 6. Compositions of feldspars in Mississippi River sand, shallow Pleistocene sands (<1.5 km of burial), and deep Pliocene sands (>4.5 km of burial). Number of grains plotted are indicated on each triangle.

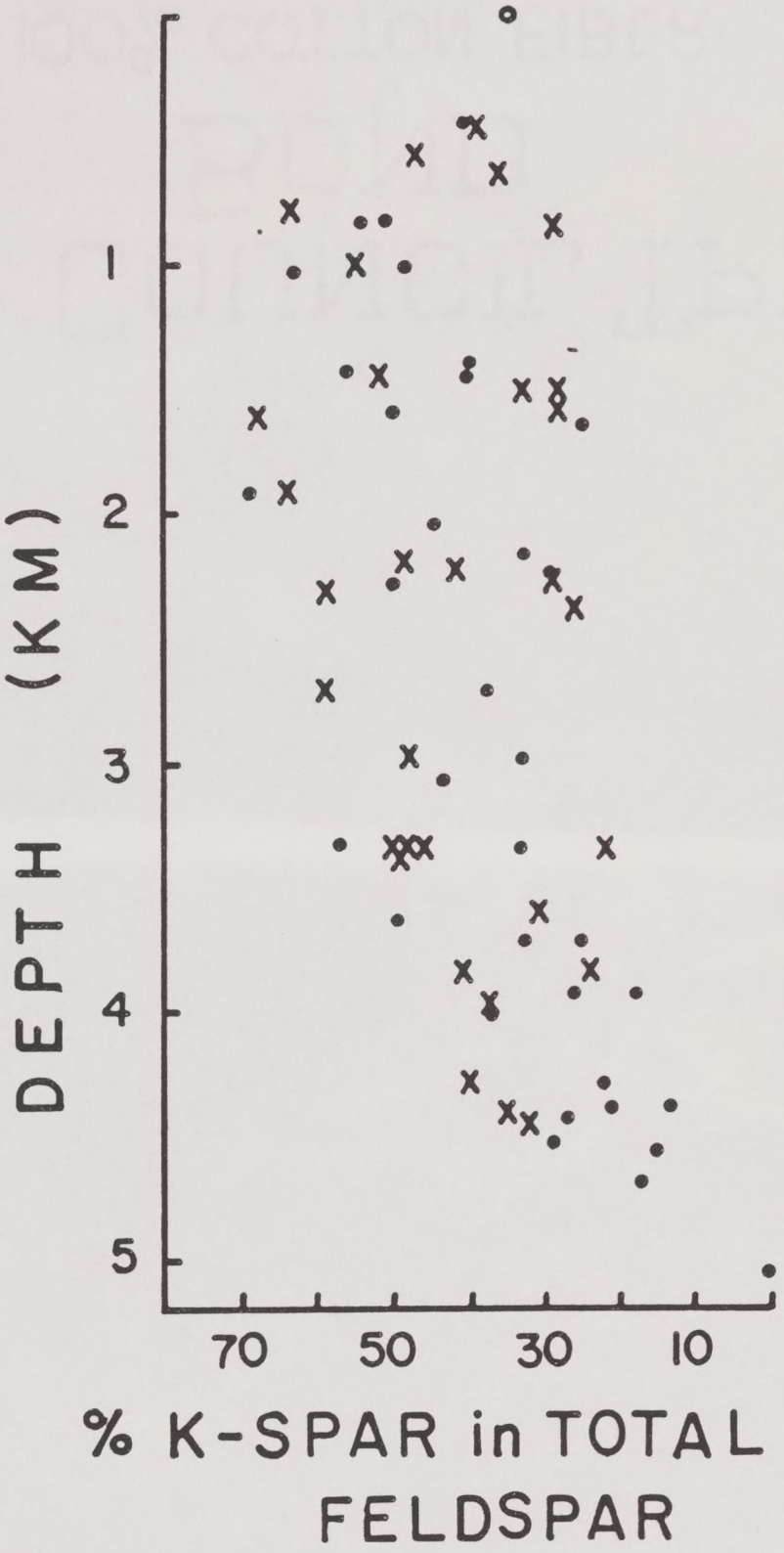


Figure 7. Percent potassium feldspar in total feldspar versus depth. Based on point counts of the >62 um fraction on the electron microprobe (●) and on thin section petrography (x). Open circle represents a sample of Mississippi River sand.

Figure 8. Potassium feldspar diagenesis. A. Small overgrowths of potassium feldspar (PP-787a, depth=2455 m). B. Dissolution of potassium feldspar along cleavages (PP-449, depth = 569 m).

C

Figure 8 continued. C. Random, "karst-like" potassium feldspar dissolution (PP-159, depth=4376 m).

some grains even have small overgrowths of authigenic potassium feldspar (Figure 8A).

Abundance of potassium feldspar relative to plagioclase diminishes with decreasing grain size (Figure 9). Grain size generally decreases with depth in the Plio-Pleistocene and thus, some portion of the decline in potassium feldspar content might be a function of grain size. Figure 9 however, shows that diagenesis also affects this trend. Samples above the depth of potassium feldspar decline display a clear relationship between grain size and the relative amounts of plagioclase and potassium feldspar. Deeper samples plot significantly above this trend, having less potassium feldspar than samples of equivalent grain size at shallower depth. Thus, loss of potassium feldspar through some diagenetic process is convincing. Evidence from the SEM as well as a few thin sections substantiates the idea that potassium feldspar is subject to dissolution and does not disappear through albitization or source-area effects (Figure 8, B and C). This supports a similar conclusion concerning potassium feldspar in the Oligocene Frio Formation of the central Texas coast by Land and Milliken (1981). Generally, though, the secondary porosity that might be expected from wholesale removal of potassium feldspar

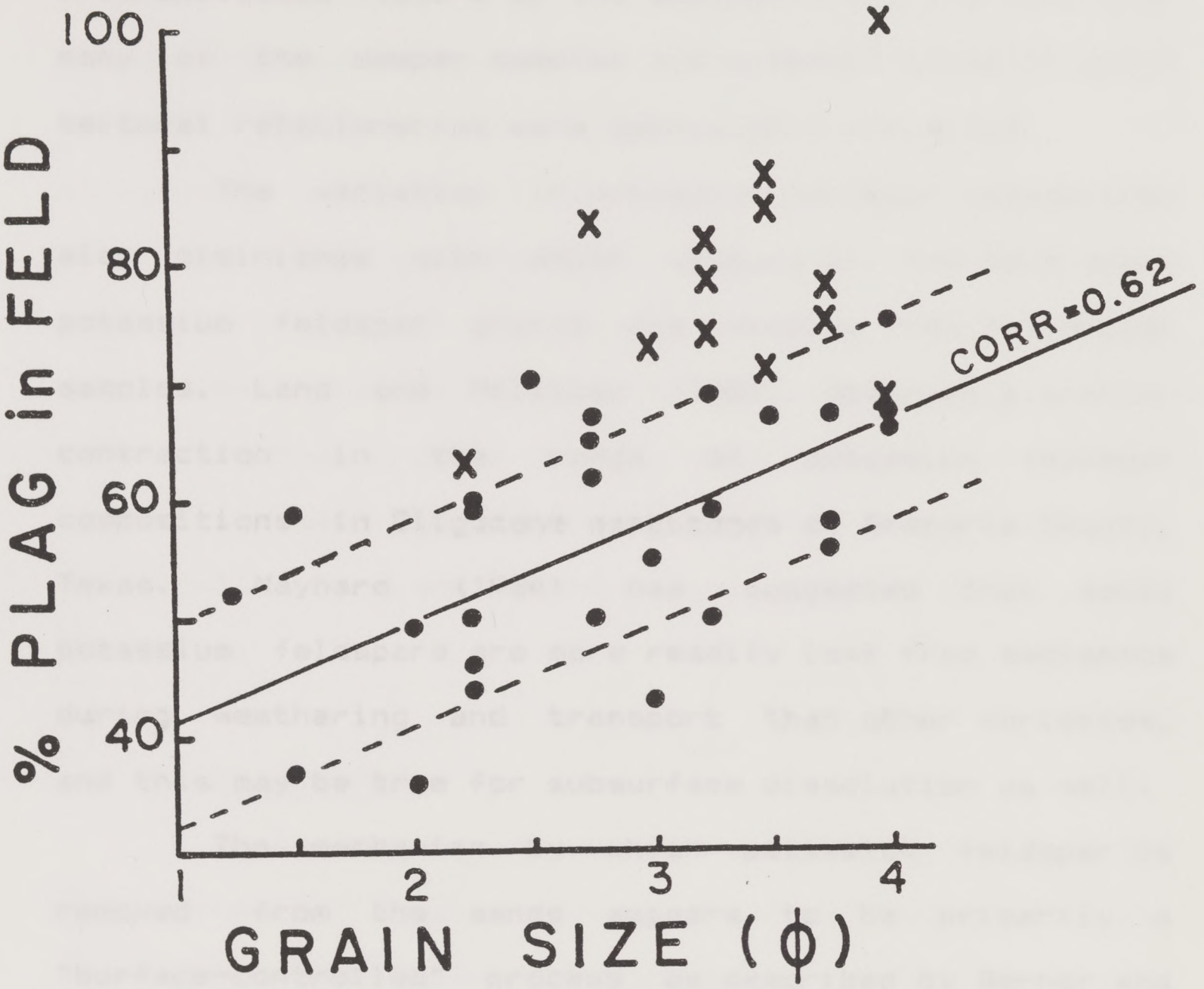


Figure 9. Percent of plagioclase in total feldspar versus grain size. Two populations of samples are apparent. Shallower than 3600 m (●) the percent plagioclase increases linearly with decreasing grain size (r=0.62; dashed lines represent ± one standard deviation). Deeper than 3600 m (x) samples are more plagioclase-rich than samples of the same grain size at shallower depth, strongly suggesting that the ratio of plagioclase and potassium feldspar has been modified through diagenesis.

was not observed in the Plio-Pleistocene. This is not especially surprising in light of the relatively unconsolidated nature of the sediments and the fact that many of the deeper samples are sidewall cores in which textural relationships were appreciably disrupted.

The variation in potassium feldspar composition also diminishes with depth (Figure 6). The more sodic potassium feldspar grains are missing from the deeper samples. Land and Milliken (1981) observed a similar contraction in the range of potassium feldspar compositions in Oligocene sandstones of Brazoria County, Texas. Maynard (1984) has suggested that sodic potassium feldspars are more readily lost from sediments during weathering and transport than other varieties, and this may be true for subsurface dissolution as well.

The mechanism by which potassium feldspar is removed from the sands appears to be primarily a "surface-controlled" process as described by Berner and Holdren (1979). This is evidenced by the strong tendency for dissolution to be localized along cleavages and/or other crystallographically-controlled features (Figure 8B). In a few of the deepest samples, however, there is evidence that another mechanism, perhaps best described as "transport-controlled", may also play a role (Figure 8C). In these grains, potassium feldspar dissolution

seems to be random. In the SEM, grain surfaces appear karst-like; in thin-section, grains are traversed by irregular canals. The appearance of this type of dissolution texture near the depth of complete potassium feldspar removal is likely a reflection of the increasing instability of the detrital grains in this zone. It would seem that at some point, the tendency of potassium feldspar to dissolve becomes sufficiently great that relative energy differences between different areas of the crystal lattice become negligible and the crystal dissolves randomly, at a rate controlled principally by transport of dissolved material away from the grain.

Plagioclase

Microprobe analyses of plagioclase (Figure 10) also reveal a depth-related trend, albeit more subtle than the trend seen for potassium feldspar. Calcium contents, averaging around An 25 in samples shallower than 3300 m, are greater than unalbitized feldspar in both Eocene and Oligocene units of the Texas Gulf Coast (Land, 1984b; Land and Fisher, in prep.). Below 3300 m the lower An content reflects both loss of the more calcic plagioclase (Figure 10C) and the addition of substantial amounts of new albite (Figure 10B).

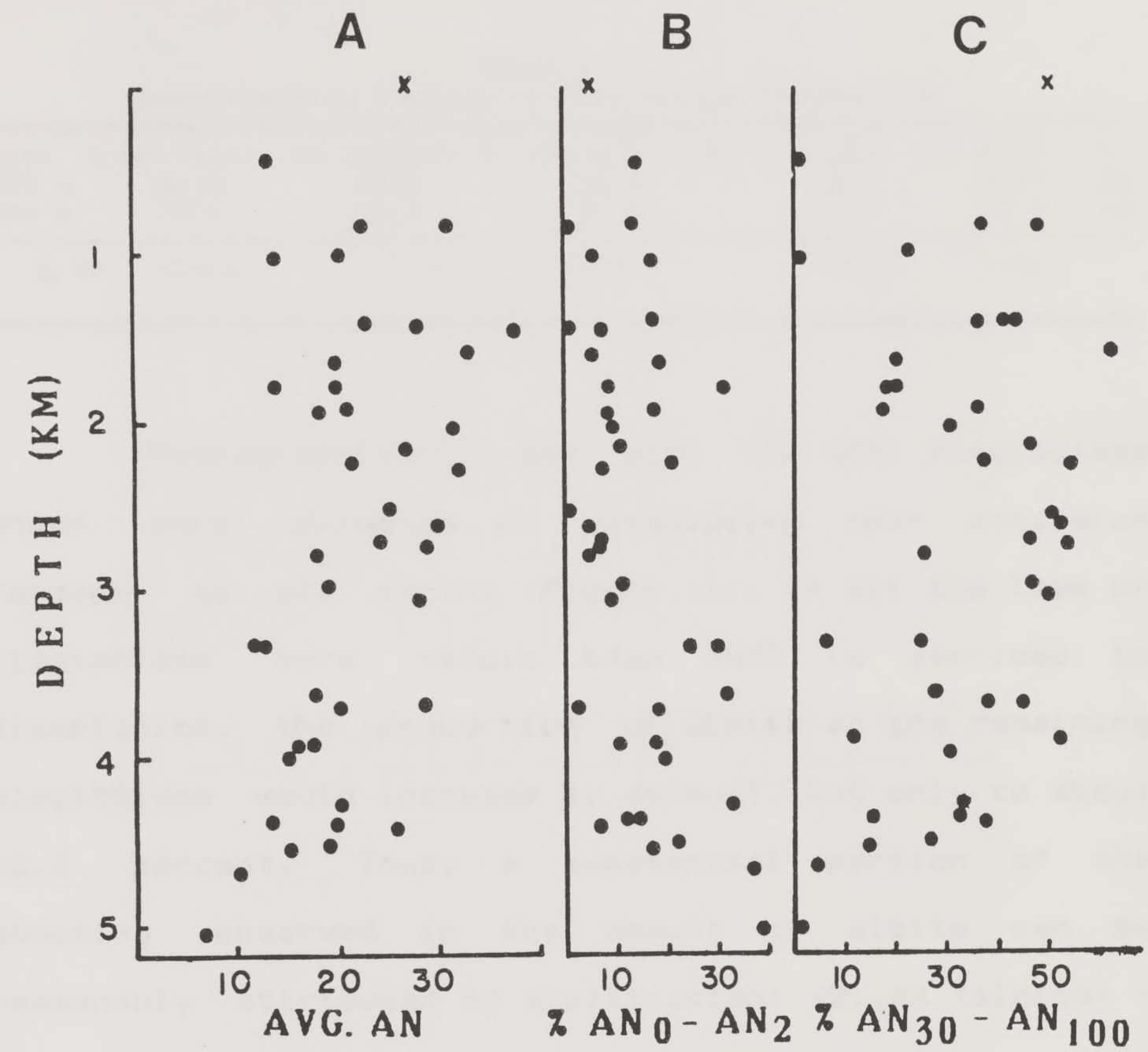


Figure 10. Composition of plagioclase versus depth. A. Average An content of plagioclase in the > 62 u fraction. B. Percent of total plagioclase that is less calcic than An 2. C. Percent of total plagioclase that is more calcic than An 30. Shallowest sample (x) is sand from the modern Mississippi River.

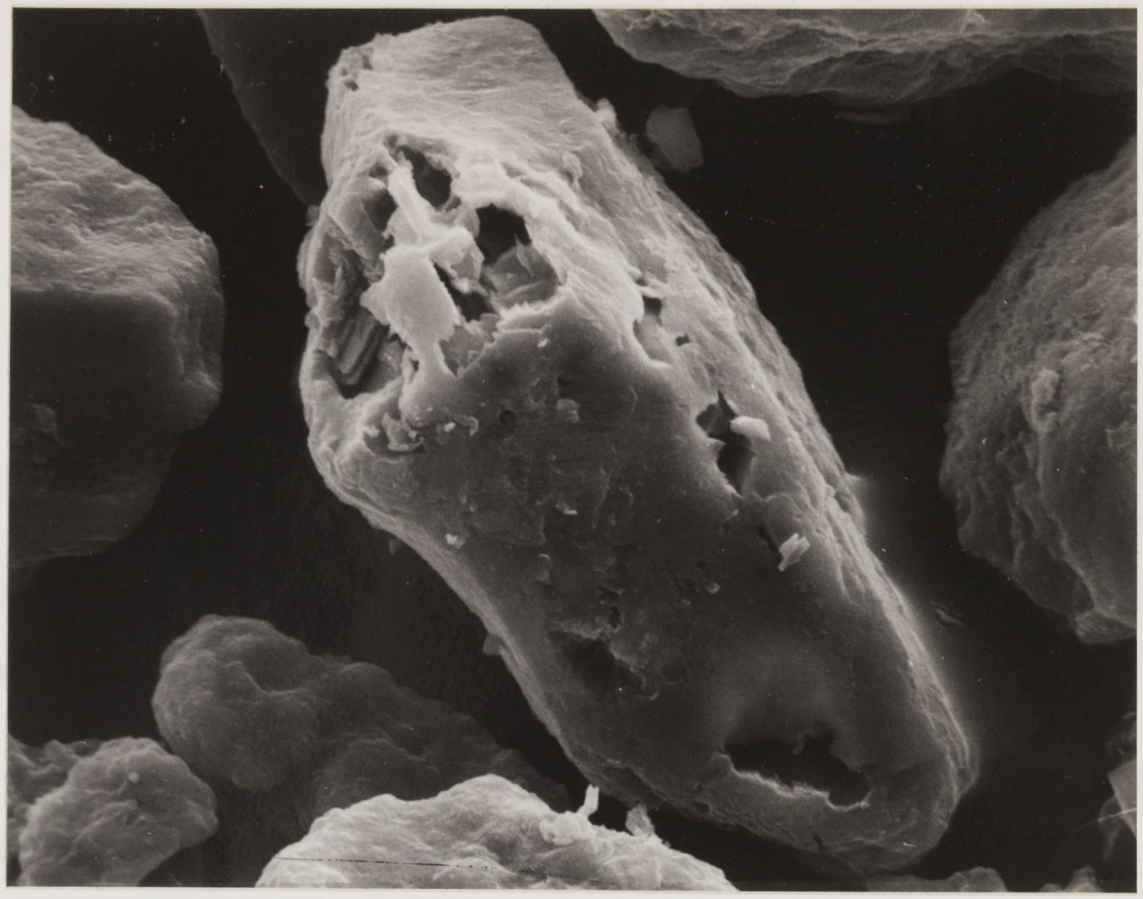
Table 3 summarizes the depth-related changes in the plagioclase population.

TABLE 3
Depth-related Changes in Plagioclase Composition

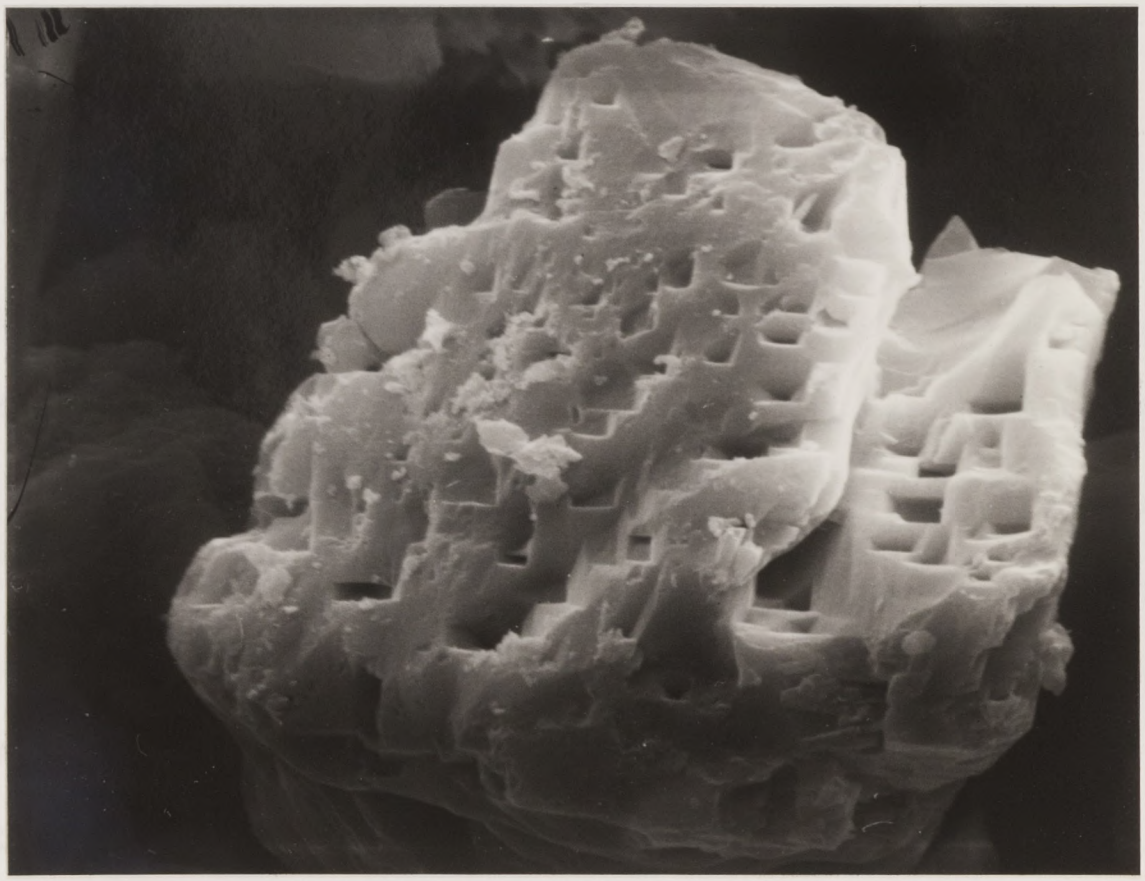
Depth	% of Feld.	AN content	An content (-Ab)	% < AN2	% > AN30	n
<3300 m	56.0	24.1	26.6	10.6	35.9	26
>3300 m	70.6	16.7	20.8	21.4	13.1	16
△ %=	+14.6	-7.4	-5.8	+10.8	-13.1	

Petrographically and with the SEM, plagioclase shows more evidence of dissolution than potassium feldspar at all depths (Figure 11). If all the loss of plagioclase more calcic than AN30 is ascribed to dissolution, the proportion of albite in the remaining plagioclase would increase by default, but only to about 12.2 percent. Thus, a substantial portion of the doubling observed in the amount of albite can be reasonably attributed to albitization: (21.4% (albite) - 12.2 % (original detrital albite, maximum) = 9.2% (albitized). Based on the composition of unalbitized plagioclase (exclusive of initial detrital albite, Table 4), and an average of 10.7 percent feldspar in the whole rock, the amount of sodium required for the increase in albite is 13 moles/m^3:

(10.7 % feld) x (56 % plag in feld) x (9 % of plag albitized) = 0.5 % of rock albitized



A



B

Figure 11. Plagioclase alteration. A. Minor dissolution of plagioclase, SEM view (PP-44, depth < 2616 m). B. Etch-pits in plagioclase similar to "surface-controlled" dissolution features described by Berner and Holdren (1979) (PP-26, depth < 1796 m).

0.005 cm3 x 2.66 g/cm3 x 1/(266.3 g/mole) = 5 x 10-5 moles An26 = 1.3 x 10-5 moles An100 = 13 moles/m^3 Ca replaced by Na in each m^3 of sand

A cubic meter of sand with 10 percent porosity which contains pore fluid of sea water salinity would have approximately 47 moles of sodium. Thus, only a few pore volumes of fluid would be required to accomplish the degree of albitization observed. More extensive albitization, as observed in Oligocene rocks of the Texas Gulf coast, would require sodium from some additional source, such as salt dissolution (Land and Milliken, 1981).

Rock fragments

Rock fragment portions are subequally divided between microcrystalline quartz and SRFs, low-rank MRFs, and acidic to intermediate VRFs (Figure 12, Figure 13). Among grains of microcrystalline quartz, some are clearly of sedimentary origin (Figure 13A), others are clearly silicified volcanics, and some are of uncertain derivation. Alteration of rock fragments beyond a minor amount of secondary dissolution was not observed, though any feldspar contained in rock fragments would be subject to the same types of alteration discussed in the section on feldspars.

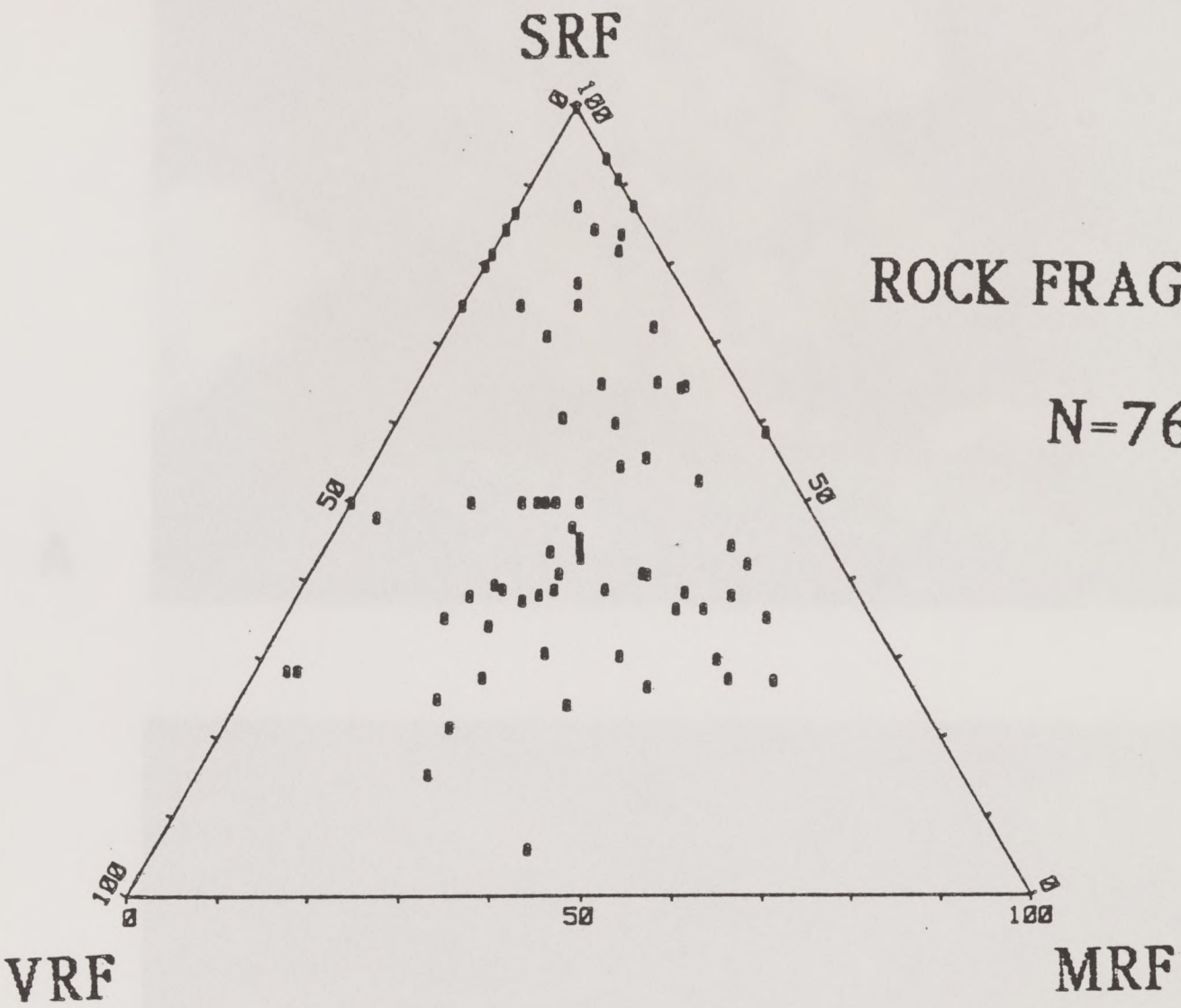
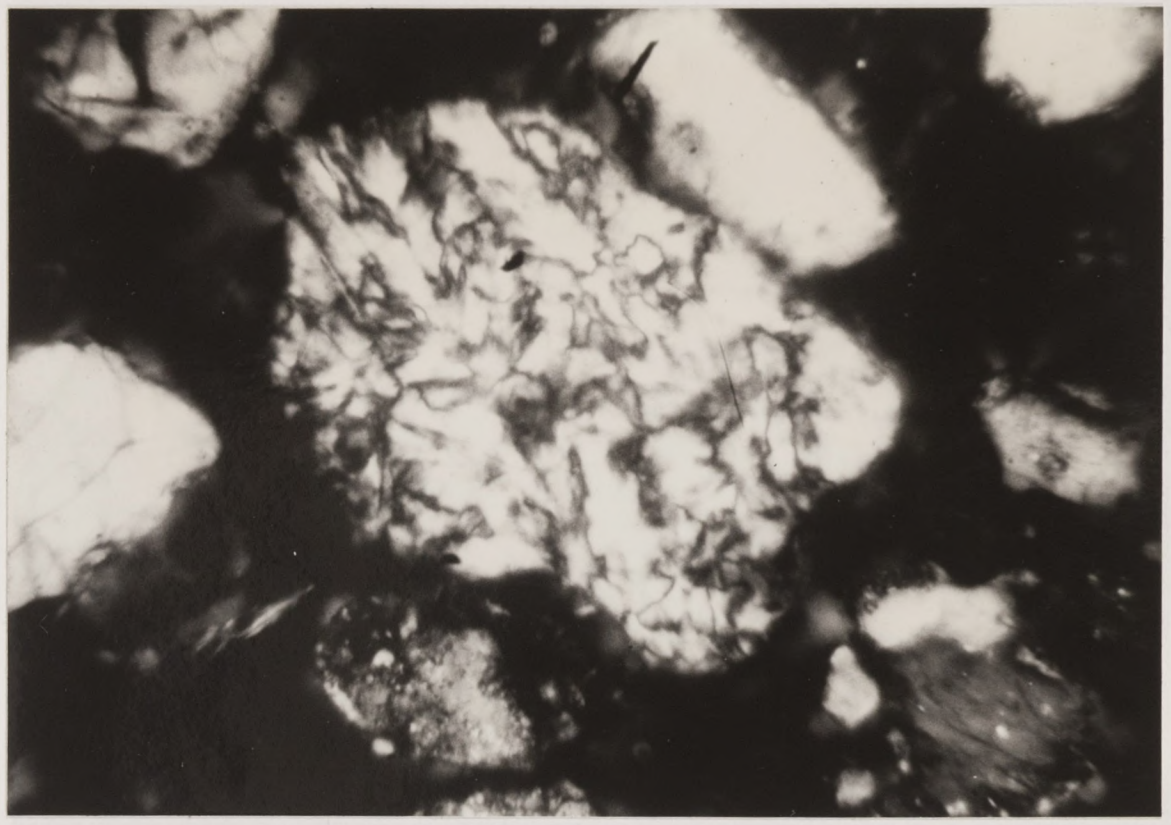
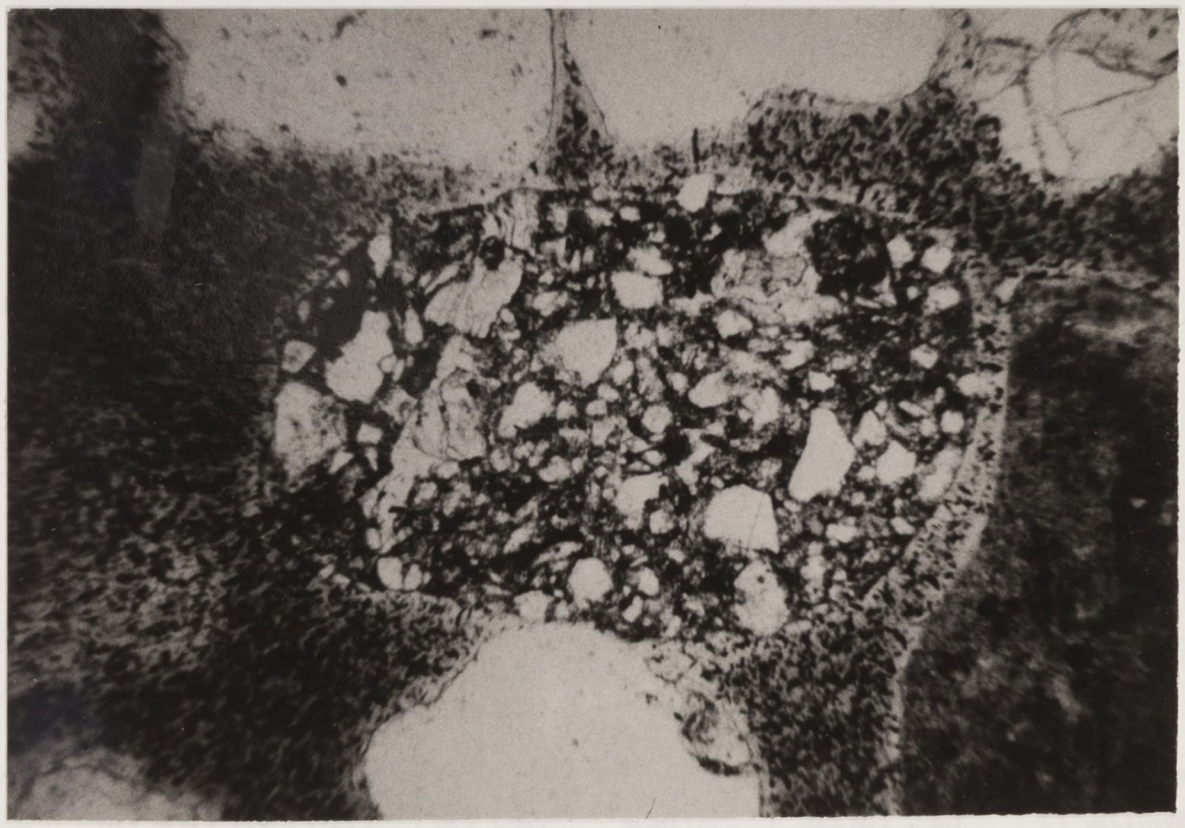


Figure 12. Rock fragment triangle. Sedimentary fragments such as mudstone and siltstone fragments are combined with microcrystalline quartz fragments at the SRF pole. Some unidentified portion of the micro-quartz may be derived from devitrified volcanics, but texturally these are difficult to discriminate from true sedimentary chert. The MRF pole includes metamorphic rock fragments, both low-rank and high-rank, as well as some possible plutonic rock fragments, mostly aggregates of equant quartz and feldspar crystals. The large variability in the triangle is due in part to grain size variations and in part to localized abundance of mud clasts likely formed as rip-up clasts.



A



B

Figure 13. Rock fragments. A. Microflamboyant quartz typical of silicified evaporites that are common in the Mississippi River drainage basin (e.g. Milliken, 1979) (PP-118, depth=2229 m). Crossed polars. B. Siltstone fragment (PP-439, depth=443 m). Plane polarized light.

0.1 mm

C

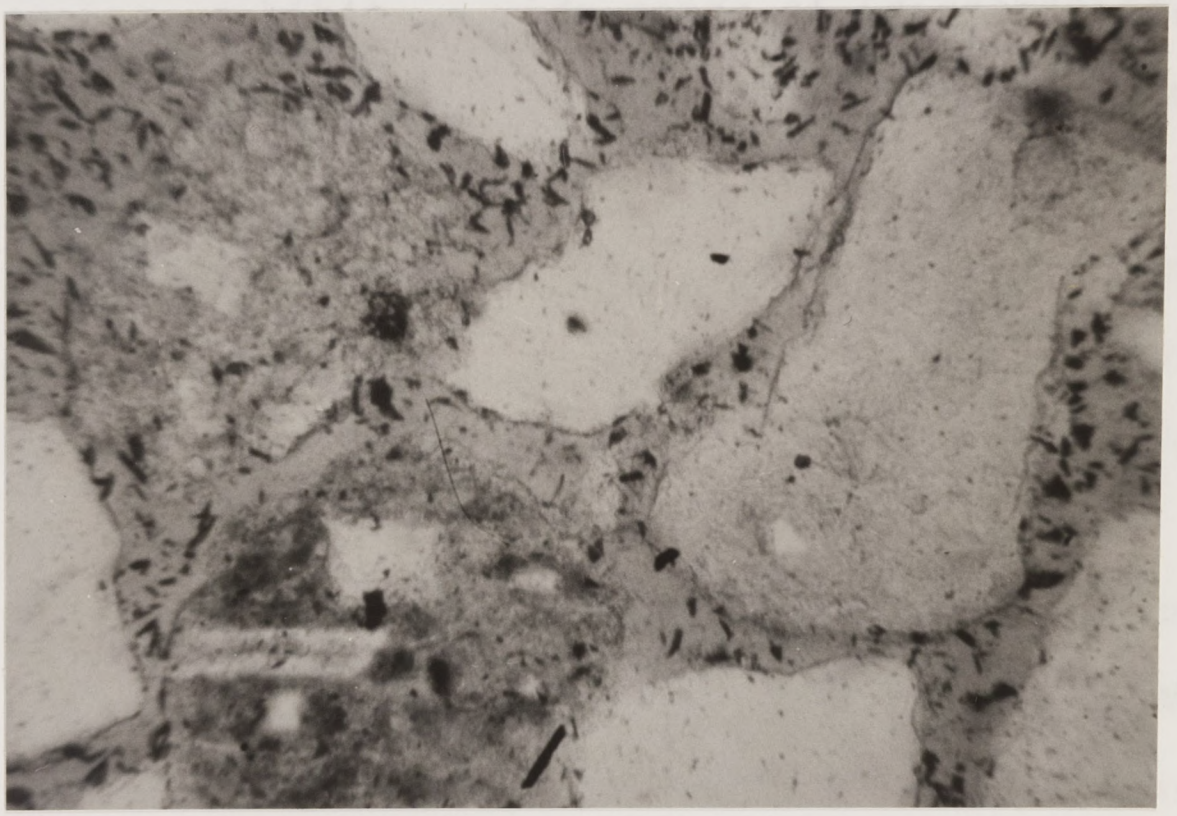
0.1 mm

D

Figure 13 continued. C. Muscovite phyllite fragment from Mississippi River sand. Similar fragments are also observed in the Plio-Pleistocene. Crossed polars. D. Volcanic rock fragment. A common type of VRF in Plio-Pleistocene sands, this grain is made of plagioclase laths in a matrix which stains with sodium cobaltinitrite, suggesting the presence of fine-grained potassium feldspar(PP-772, depth=2738 m). Plane polarized light.

0.05 mm

E



0.1 mm

F

Figure 13 contiunued. E. VRF made of plagioclase laths with a highly altered groundmass. Alteration appears to be primarily iron oxides and/or sulfides, suggesting the former presence of ferromagnesian minerals (PP-81, depth=2272 m). Plane polarized light. F. Highly clay-altered VRFs typical of Plio-Pleistocene sands (PP-74, depth=2036 m). Plane polarized light.

Included among SRFs are a number of carbonate crystals, typically mono-crystalline dolomite, which are interpreted as detritus from carbonate rocks (Figure 14). An alternative, of course, would be to interpret these crystals as grain replacements, features which are commonly observed in older Tertiary rocks of the Gulf Coast (Loucks and others, 1979). However, a detrital origin seems more likely for the carbonate grains in the Plio-Pleistocene for the following reasons:

1. The crystals are not particularly euhedral. Within a particular sample the grains exhibit a wide range of idiomorphism, suggesting they have been subject to abrasion.
2. No examples of partially replaced grains were observed. In the older Tertiary rocks of the Texas coast, partial replacement is common and convincing.
3. Mississippi River sand contains carbonate grains, clearly detrital, which resemble the carbonate grains in the Plio-Pleistocene in size, petrographic appearance, and abundance.
4. Stable isotopic analysis of carbonate in the Mississippi River sand ($\delta^{18}O$=-4.7 o/oo PDB, $\delta^{13}C$=-1.7 o/oo PDB) is consistent for

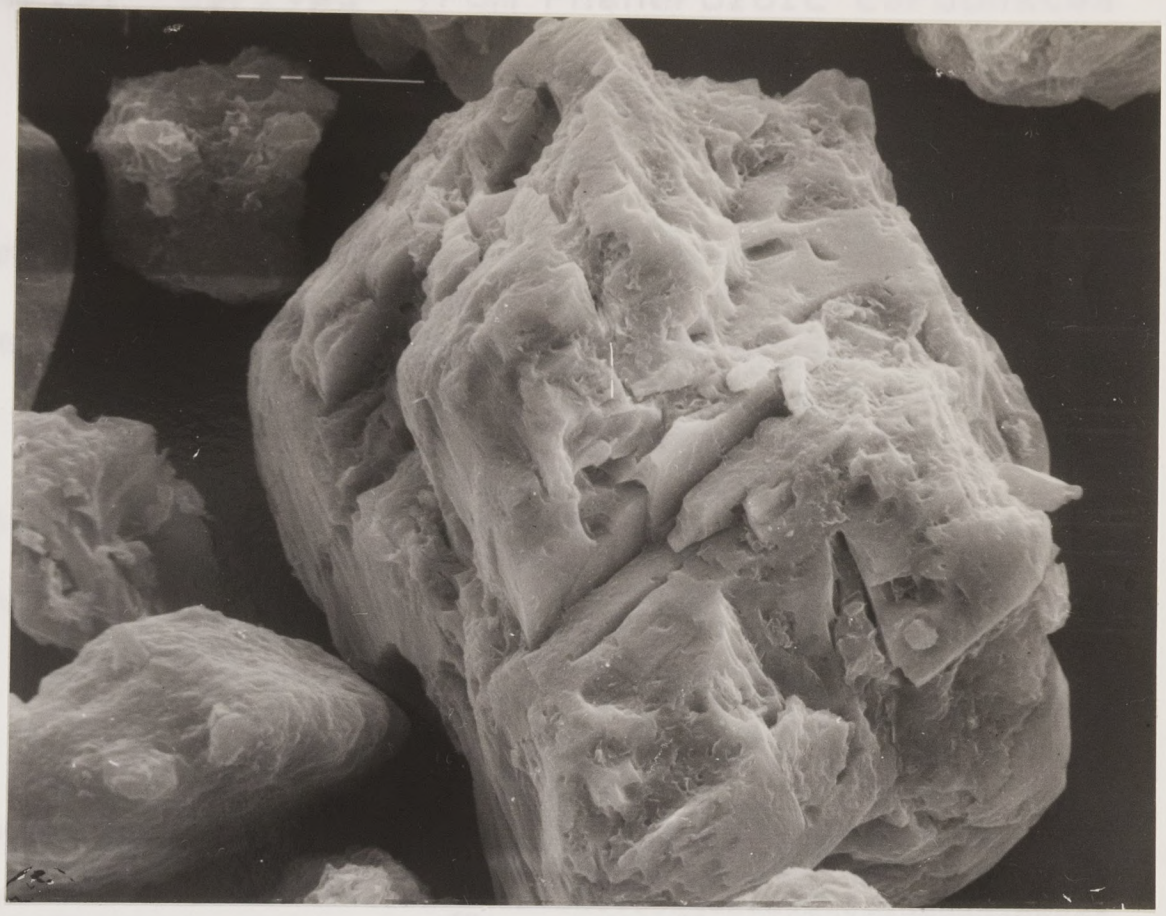
A

0.05 mm

B

0.05 mm

Figure 14. Carbonate crystals interpreted as detrital grains. A. Carbonate grain (probably dolomite) with a small overgrowth (probably siderite) (PP-770, depth=2737 m). Plane polarized light. B. Dolomite grain from Mississippi River sand. Plane polarized light.



C

Figure 14 continued. C. SEM view of abraded and pitted carbonate grain (PP-12, depth < 1780 m).

material derived from Phanerozoic carbonates of the mid-continent. Carbonate samples from the Plio-Pleistocene which are comprised primarily of the detrital-looking dolomite crystals have oxygen values similar to that of carbonate in Mississippi River sand, though carbon values tend to be slightly heavier (Table 4).

TABLE 4
Carbon and Oxygen Isotopic Composition of Dolomite Grains in Sandstones

Number	Depth (m)	$\delta^{18}O$ (PDB)#	$\delta^{13}C$ (PDB)#	Weight %*
Miss. R.	0	-4.7	-1.6	0.7
PP-12	1780 (md)	-4.0	-0.4	3.7
PP-13	1782 (md)	-5.0	+0.1	3.6
PP-18	1785 (md)	-4.5	-0.3	3.7
PP-66	1629	-4.7	-0.7	---
PP-288	5046	-5.2	-1.9	3.0

#±0.1 o/oo
* Determined manometrically

Minor detrital constituents

The most abundant minor constituents are skeletal debris, opaque and non-opaque heavy minerals, and glauconite. The average amount of these minor constituents in the consolidated samples reported in Appendix 1 is 1.2 percent of the total rock volume.

Skeletal debris

Invertebrate skeletal debris found in the sands includes pelecypods, gastropods, echinoids, foraminifers (benthic and pelagic), corals (Oculina sp.), and serpulid worm tubes. Invertebrate assemblages on the modern shelf contain similar organisms (e.g. Parker, 1960).

Several interesting examples of pressure solution of skeletal debris were observed (Figure 15). Some of these examples are from quite shallow depths and it is possible pressure solution may be important in removal of skeletal debris from the sediment and to early recycling of this material into authigenic phases.

Heavy minerals

Opaque heavy minerals present include magnetite, ilmenite, pyrite (mostly authigenic?), and leucoxene. No attempt was made to estimate the relative percentage of these minerals or their possible variations with depth. No evidence of opaque mineral dissolution was observed in thin section, although etching of these grains was seen with the SEM.

Non-opaque heavy minerals show an interesting variation with depth (Figure 16). Heavy mineral assemblages in Gulf Coast rivers and on the modern shelf

0.2 mm

Figure 15. Pressure solution of an oyster fragment, penetrated by quartz grains (PP-172, depth=2971 m). Plane polarized light.

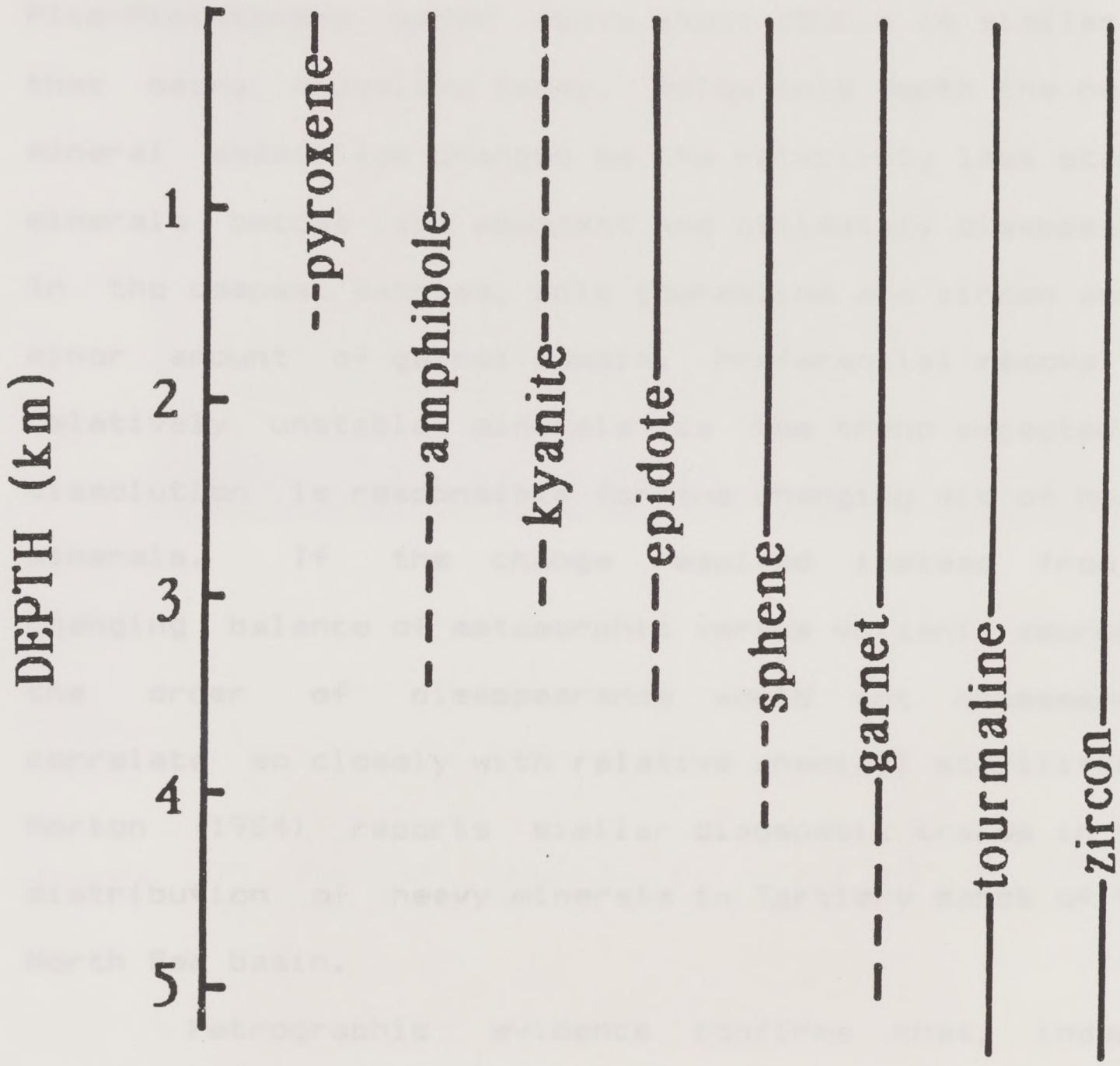
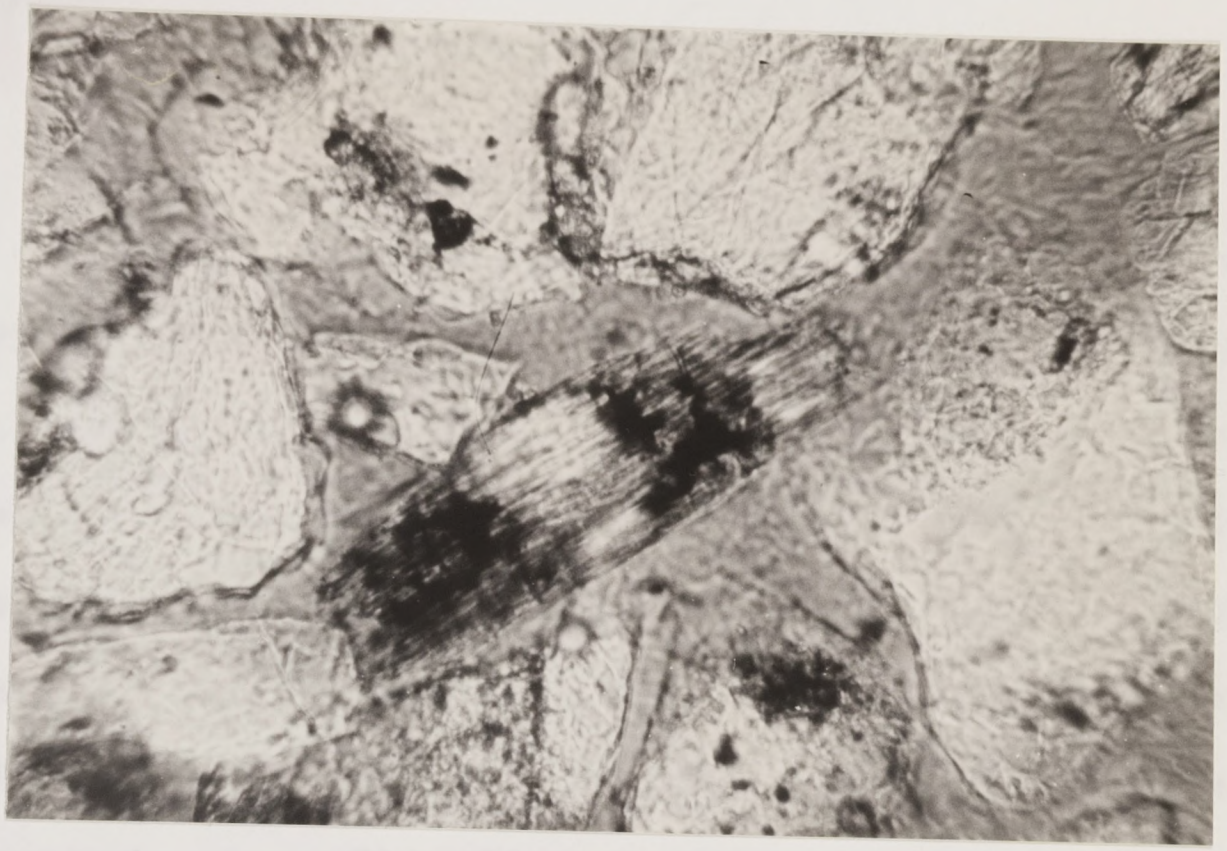


Figure 16. Occurrence of heavy minerals in Plio-Pleistocene sediments. Based simply on presence/absence as observed in thin section. Lines are solid where a mineral is observed in most samples and dashed where the mineral is observed in only a few samples. Staurolite distribution is similar to that of kyanite.

are fairly well documented (e.g. Van Andel, 1960; Davies and Moore, 1970). Based on comparison with this published data, the heavy mineral assemblage in Plio-Pleistocene sands above about 1500 m is similar to that being deposited today. Below this depth the heavy mineral assemblage changes as the relatively less stable minerals become less abundant and ultimately disappear. In the deepest samples, only tourmaline and zircon and a minor amount of garnet remain. Preferential removal of relatively unstable minerals is the trend expected if dissolution is responsible for the changing mix of heavy minerals. If the change resulted instead from a changing balance of metamorphic versus volcanic sources, the order of disappearance would not necessarily correlate so closely with relative chemical stabilities. Morton (1984) reports similar diagenetic trends in the distribution of heavy minerals in Tertiary sands of the North Sea basin.

Petrographic evidence confirms that, indeed, non-opaque heavy minerals are subject to dissolution (Figure 17). Authigenic phases such as chlorite or kaolinite were not observed in the vicinities of leached heavy minerals (nor around leached feldspars). It seems likely, then, that the dissolution mechanism was a rather open-system process, at least on the scale



0.05 mm

Figure 17. Partial dissolution of hornblende (PP-770, depth=2737 m). Plane polarized light.

of a few centimenters, and probably over even greater distances.

Source area implications

The location and orientation of the major Plio-Pleistocene depocenters suggest that the source of these sediments was more or less equivalent with the drainage of the modern Mississippi River (Winker, 1981; Figure 18). Source rocks contributing sediment to the Mississippi have the approximate areal proportions shown in Figure 19. Each of these source terrains is represented in Plio-Pleistocene sands, though not necessarily in amounts proportional to their areas. Sedimentary rocks cover a vastly greater proportion of the area than igneous and metamorphic terrains. Judging from the abundance of calcic plagioclase, sodic potassium feldspar, volcanic rock fragments, and unstable heavy minerals, detritus from non-sedimentary sources is over-represented. This apparent over-abundance may arise because much of this unstable material is actually recycled. For instance, both the large expanse of early Tertiary sediment shed from the Rocky Mountains and the northern third of the basin covered by glacial sediment may have contributed significant amounts of unstable detritus. Chatter-mark

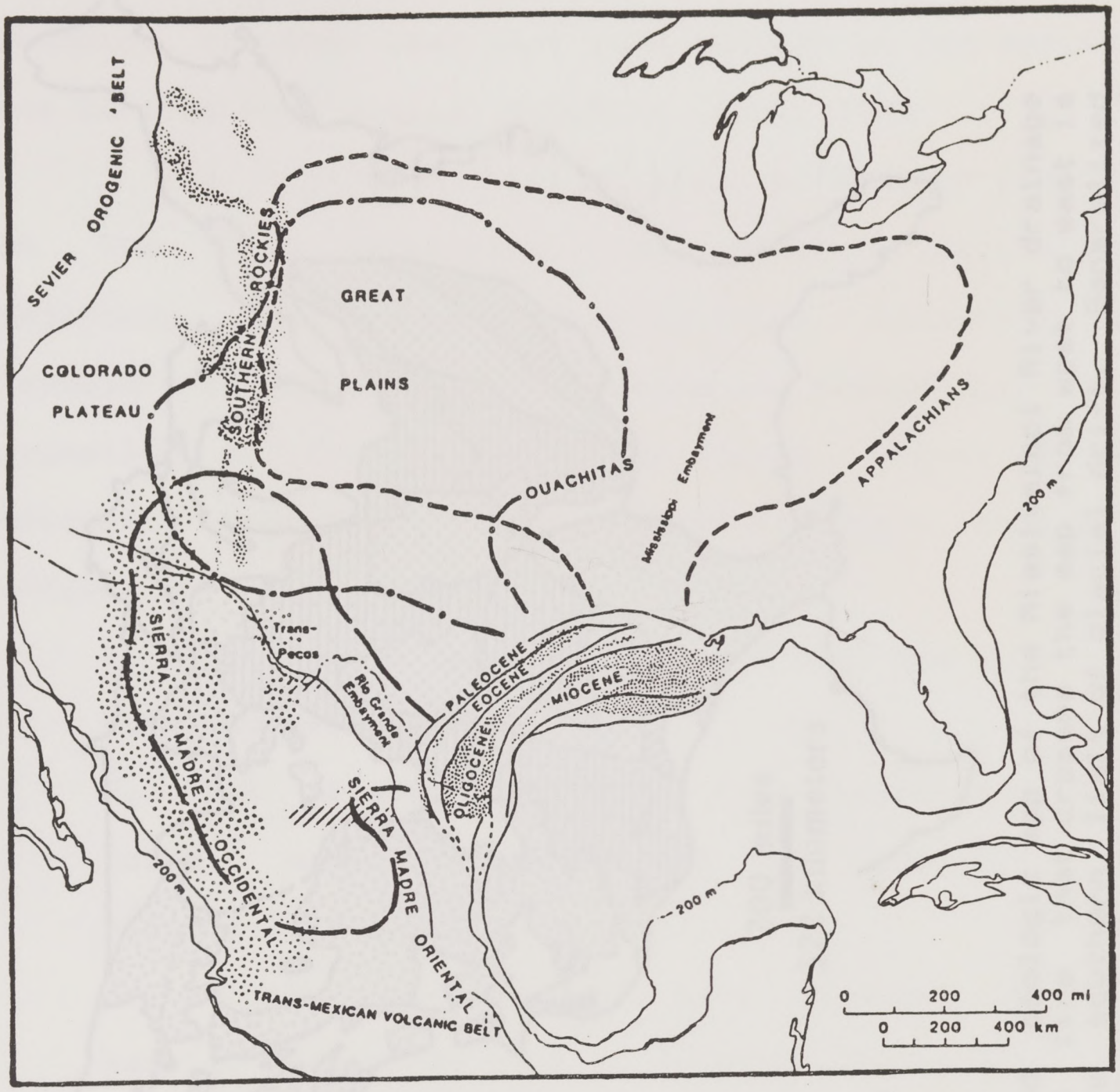


Figure 18. Inferred source areas for Cenozoic sediments in the Gulf of Mexico basin. (from Winker, 1981). Stippled areas along the coast represent the major depocenters.

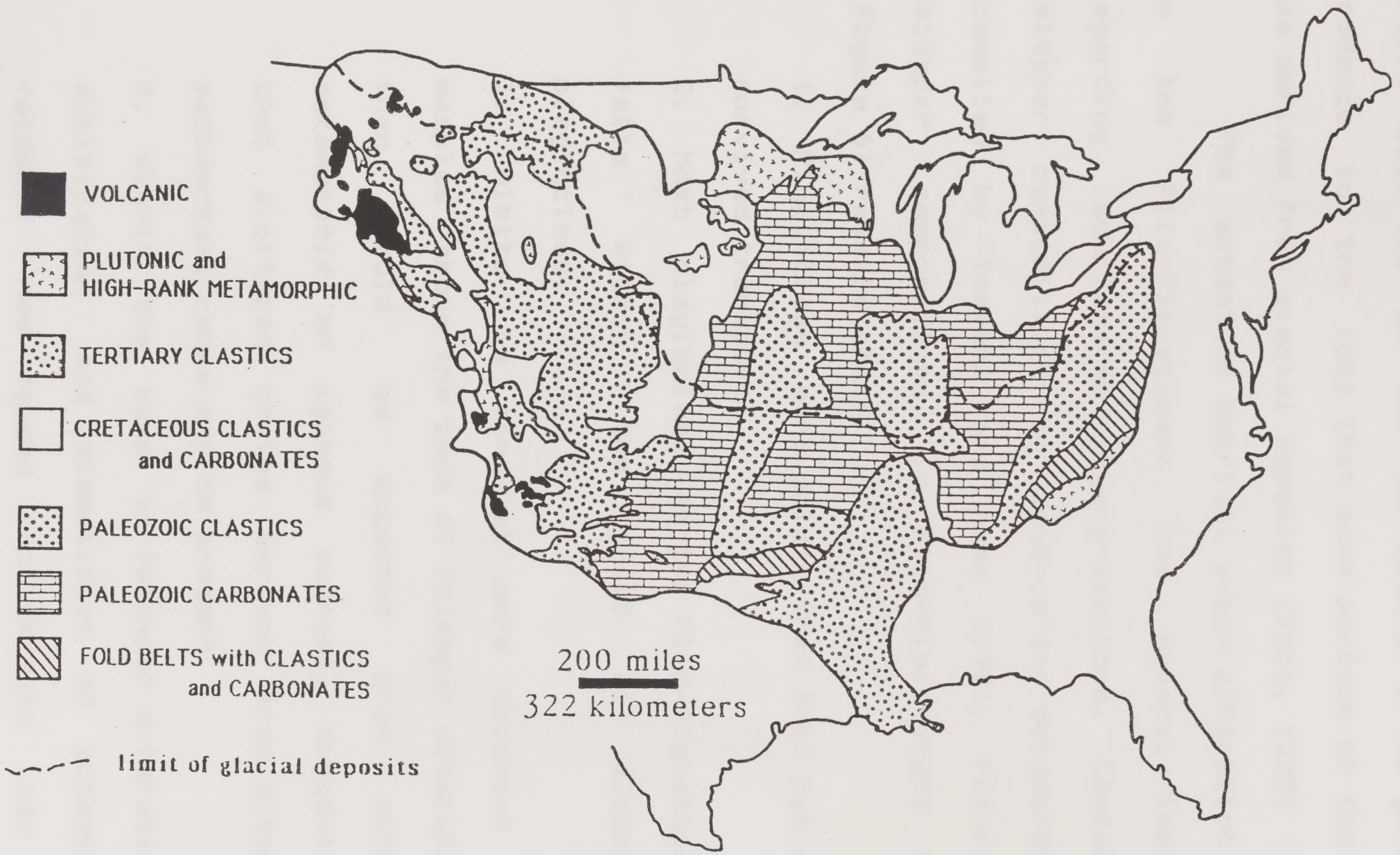


Figure 19. Generalized geologic map of the Mississippi River drainage basin. Dashed line that crosses the map from west to east is the approximate southern limit of glacial drift. Generalized from U. S. Geological Survey Geological Map of the United States 1:7,000,000 (Kinney, 1966).

trails observed on the surfaces of a few garnets give credence to the idea that some portion of the sediment was derived from glacial deposits (Folk, 1975)

The extent of detrital grain alteration observed in the Plio-Pleistocene has several implications regarding source-area interpretations. Comparison of feldspar compositions in igneous and metamorphic rocks (compiled by Trevena and Nash (1981, Figure 20) and feldspar compositions in Plio-Pleistocene sediments (Figure 6) reveals that:

1. Plagioclase more calcic than AN60 has not survived transport.

2. Most plagioclase in the Plio-Pleistocene falls within the field of plutonic plagioclase.

3. Albite is somewhat more abundant in samples above the zone of feldspar alteration than would be expected from either metamorphic or igneous sources, suggesting that albitized grains derived from altered sedimentary rocks may be present.

4. Within the zone of feldspar alteration, albitization and dissolution of potassium feldspar has caused considerable loss of information about the source. Certainly, this

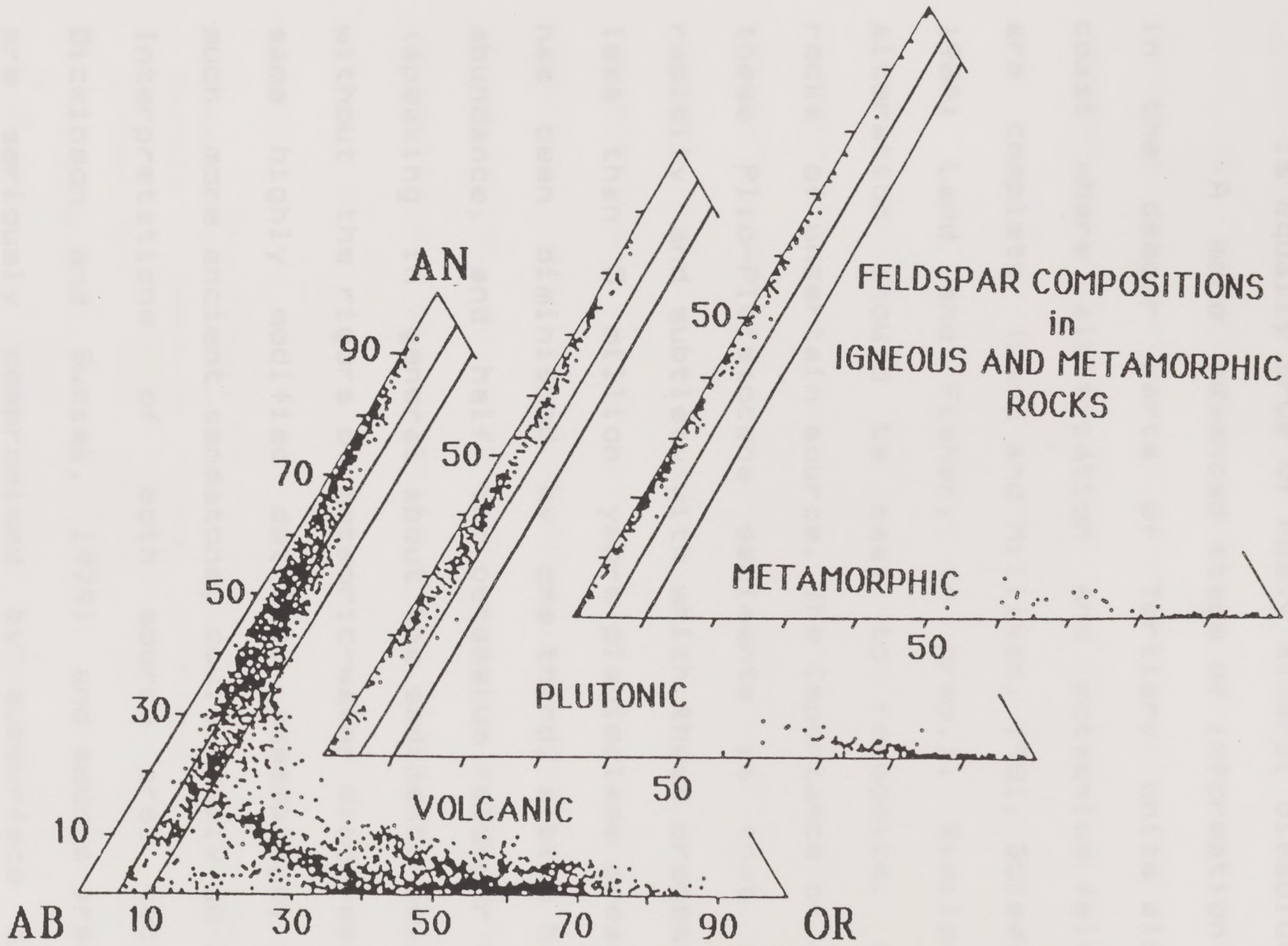


Figure 20. Compositions of feldspars in igneous and metamorphic rocks, compiled by Trevena and Nash (1981). For reference, lines have been drawn to indicate 5 and 10 percent potassium in plagioclase.

is equally true of heavy mineral dissolution.

A more advanced stage of information loss is seen in the deeper parts of Tertiary units along the Texas coast where albitization and potassium feldspar removal are complete (Land and Milliken, 1981, Boles, 1982, Land, 1984; Land and Fisher, in prep.). Similar examples of alteration should be easy to recognize, even in older rocks of uncertain source. The importance of alteration in these Plio-Pleistocene sediments is that it shows the rapidity and subtlety with which these processes begin. In less than 5 million years plagioclase greater than AN30 has been diminished by one third, albite has doubled in abundance, and half the potassium feldspar has dissolved (speaking in general about the sediments below 3 km), all without the rigors of meteoric-water diagenesis. Yet, this same highly modified detrital assemblage, if found in a much more ancient sandstone, could easily go unrecognized. Interpretations of both source area tectonics (e.g. Dickinson and Suczek, 1979) and source area lithologies are seriously compromised by subsurface alteration of detrital mineral assemblages.

Pore-filling authigenic components

Cementation of Plio-Pleistocene sediments resembles that in older Tertiary units of the Gulf Coast

in that carbonate and quartz are the dominant cements, but the degree of cementation is tremendously less. Samples of essentially unconsolidated sand are found at depths as great as 4 km in the Plio-Pleistocene.

Carbonate cements

Calcite, dolomite, and siderite, in various combinations, are all locally important cements in Plio-Pleistocene sediments. So localized, in fact, are cement occurrences that each example of significant cementation (> 2% by volume as determined in thin section) can be discussed individually. Table 5 is a compilation of data used in the following discussion; examples are discussed in order of increasing depth.

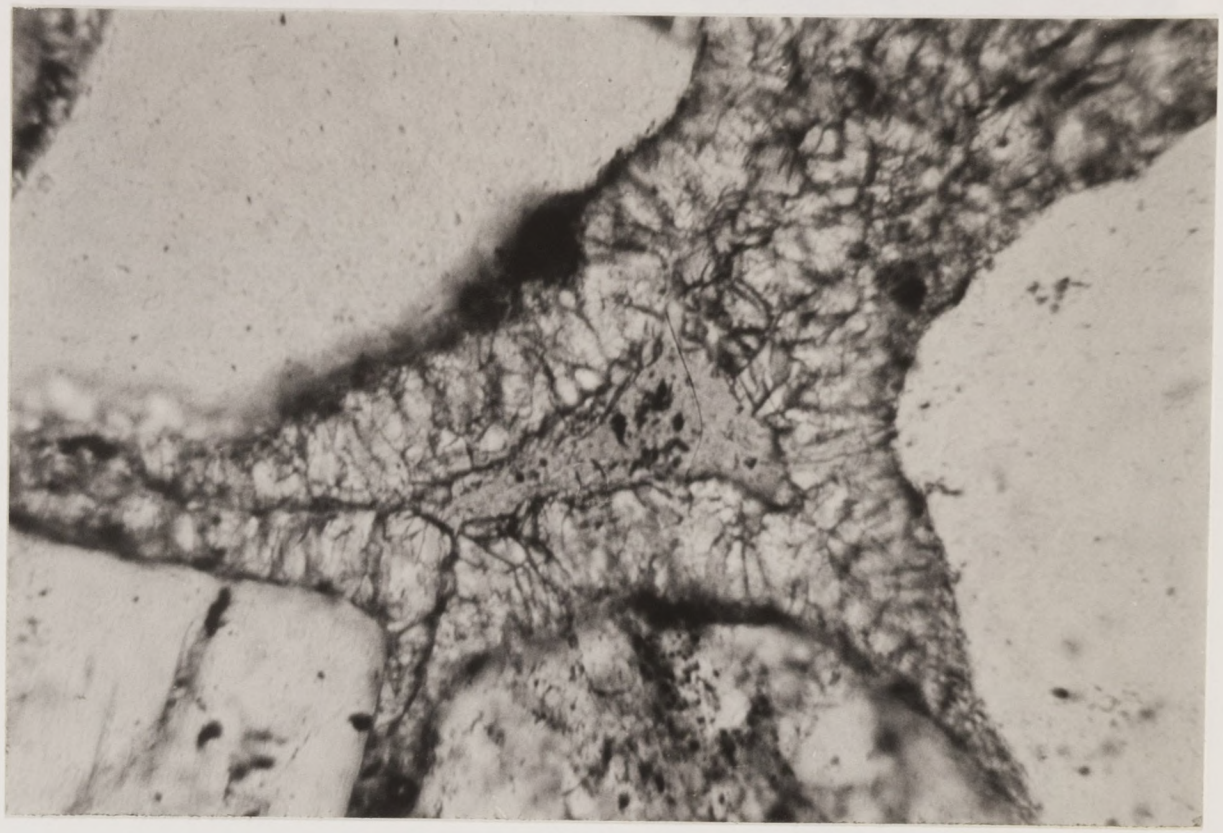
TABLE 5
Data Summary for Carbonate-cemented Sandstones*

Example	Depth (m)	Mineral	% of rock	$\delta^{18}O$	$\delta^{13}C$	$^{87}Sr/^{86}Sr$
1	<1000	dolomite	40	+4.0	-39	.7097
2 (host rock)	2200	calcite	65	-5.2	+21.4	.7084
		siderite		-1.5	+3.1	
(concretion)		siderite		-0.3	+10.0	.7089
3	2600	calcite	37	-8.4	-12.9	
4	3000	calcite?	35			
5	4500	dolomite	10	-6.8	-4.7	

*some quantities are averages of values determined on two samples

Example 1, Shallow dolomite with variable morphology: Two wells (24 and 26), drilled from a single

platform in the Eugene Island area, contain an interesting occurrence of dolomite cement, unfortunately observed only in cuttings. Cuttings with cemented sandstone in these wells are found in the depth range of 457 to 760 m (approximately). In general, the bulk of sand in this interval is unconsolidated. In thin section (PP-430, PP-449), the cement appears as coarse, crudely bladed, isopachous rims on grains (Figure 21 A) which comprise about 40 percent of the rock. In SEM, an incredible morphological diversity is revealed (Figure 21 B-D). Associated with the dolomite-cemented sandstone cuttings are pieces of dolomicrite (PP-447) containing approximately 75 percent dolomite by weight, the rest being fine-grained silicate detritus. Presumably these dolomicrites occur as thin interbeds with the dolomite-cemented sandstones. Isotopically, the dolomite in both cemented sand and dolomicrite is rather heavy with respect to oxygen (average $\delta^{18}O$=+4 o/oo) and extremely light with respect to carbon (average $\delta^{13}C$=-39 o/oo). The $^{87}Sr/^{86}Sr$ ratios of two dolomicrite samples are also somewhat unusual (Table 6) as discussed below. Depth of these samples constrains the possible range of temperature rather narrowly, so that the fluid responsible for precipitation of the dolomite must have had an oxygen composition very near 0



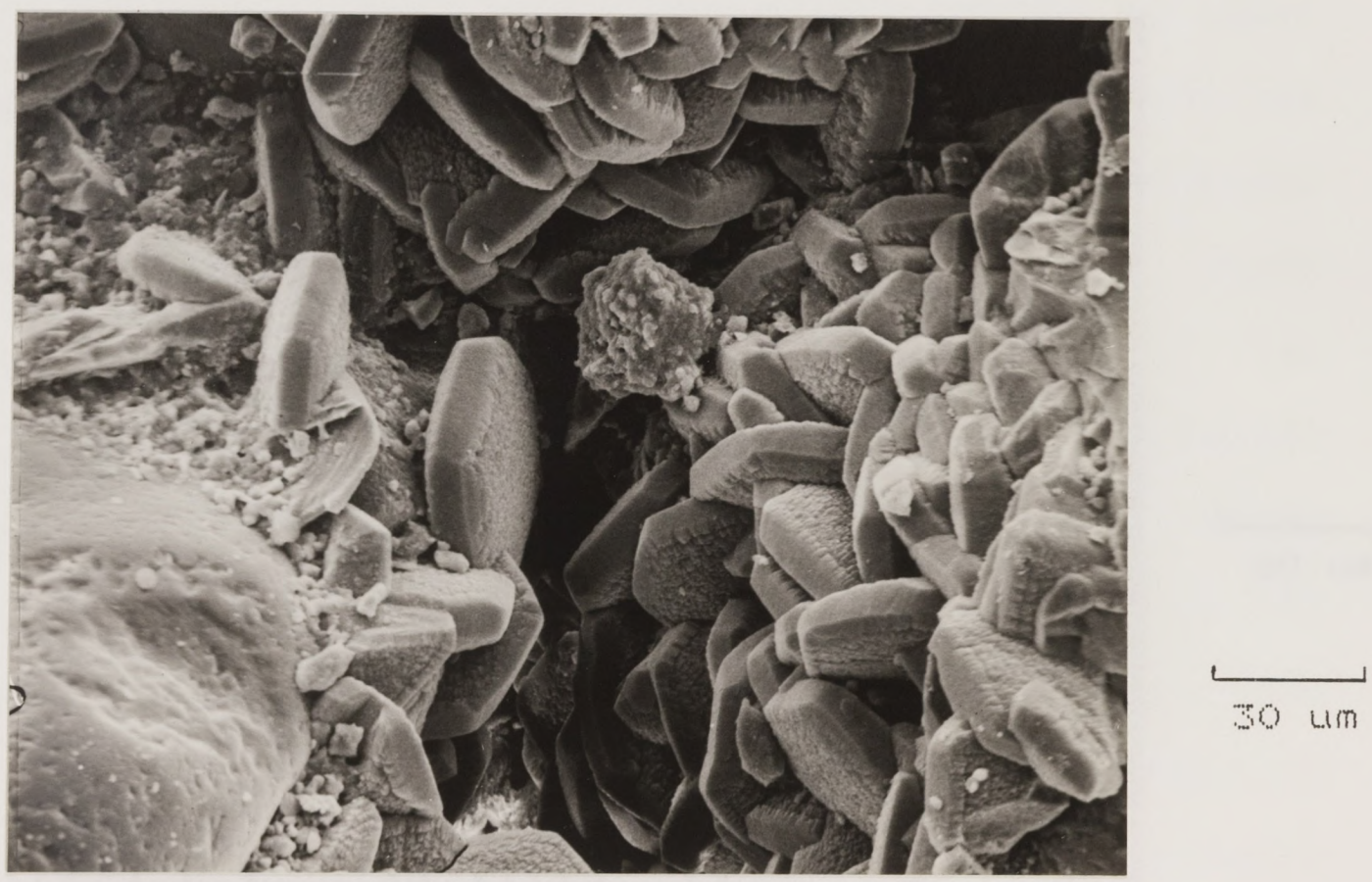
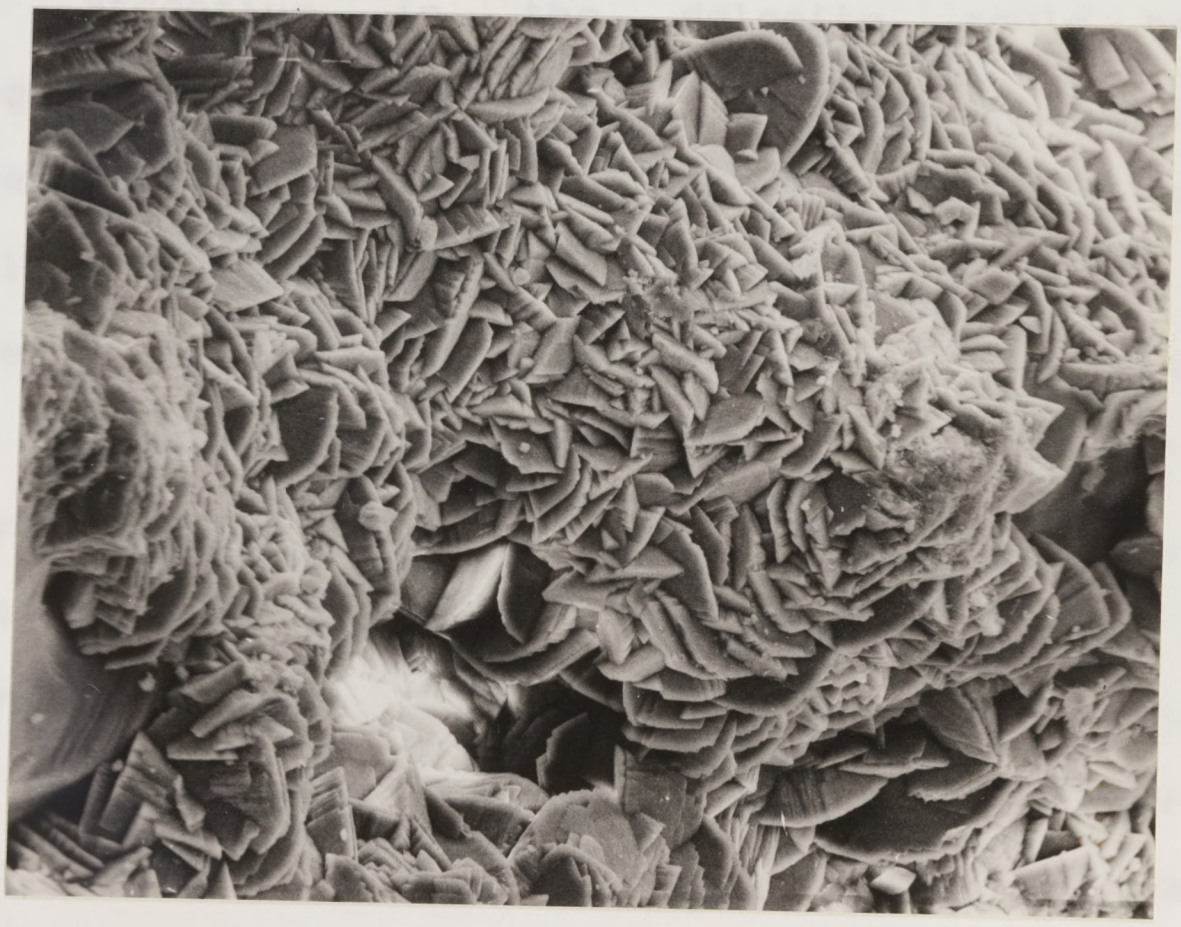


Figure 21. Shallow dolomite cements. (PP-449, depth=569 m). A. Bladed crusts as seen in thin section. Plane polarized light. B. SEM view of dolomite cement.

C



D

Figure 21 continued. C, D. SEM views of two dolomite morphologies.

o/oo. Meteoric water would require temperatures higher than the maximum possible for these samples (~40^{0} C). Extremely light carbon isotopic values have been reported for several occurrences of near-surface carbonate cementation in terrigenous marine sediments (e.g. Deuser, 1970; Russel and others, 1966; Roberts and Whelan, 1975), and these are generally ascribed to precipitation associated with oxidation of organic matter or biogenic methane. This interpretation seems entirely reasonable for this occurrence of dolomite as well, although the Sr isotopic values are somewhat difficult to explain.

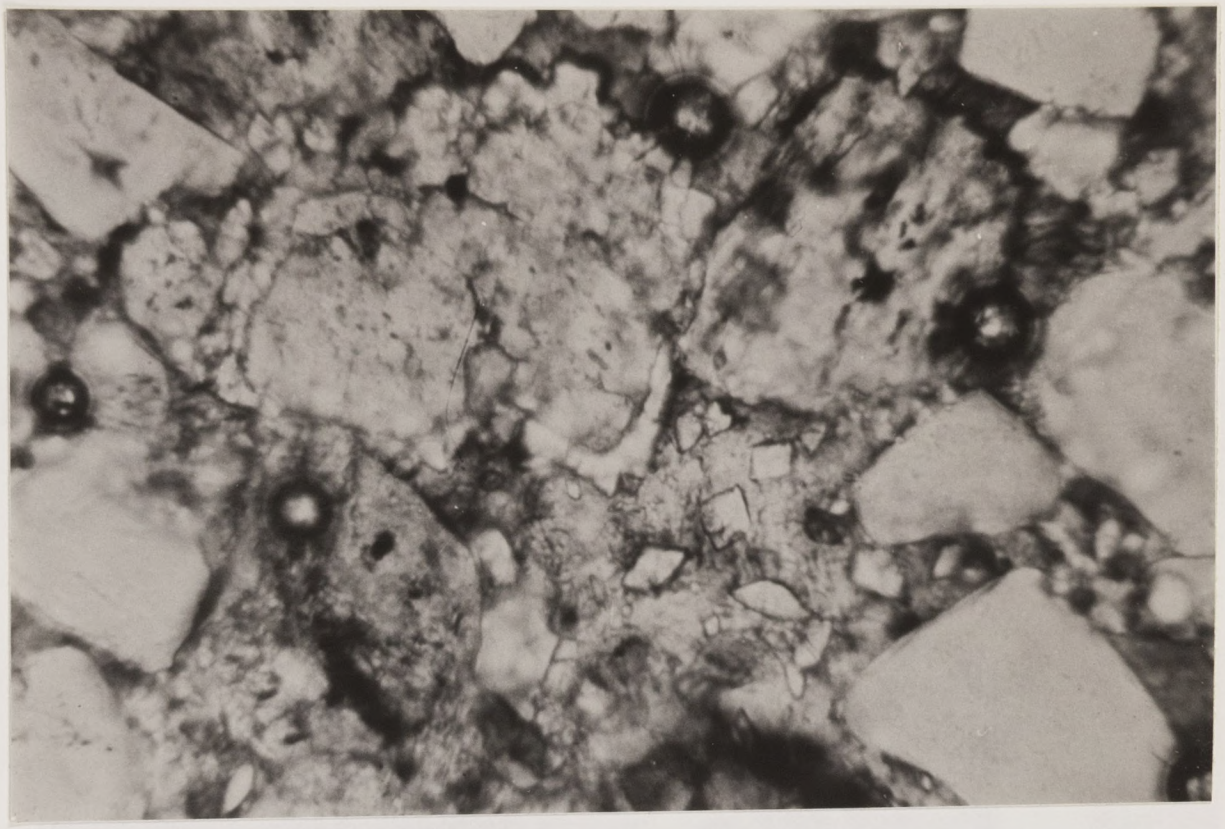
Strontium in the dolomite could be made radiogenic in at least three ways. ^{87}Sr could be delivered through: meteoric input, adsorbed Sr released by clays soon after burial, or deep basinal fluid channelled through highly localized fluid pathways. The first possibility seems clearly ruled out by oxygen isotopic composition of the dolomite. Likewise, water of deep basinal derivation would likely be of inappropriate oxygen composition unless the fluid was also warm. Release of adsorbed Sr, while an interesting possibility (suggestion of Larry Mack, 1984, personal communication), cannot be assessed without a larger data base for Sr ratios in Gulf coast pore fluids.

Example 2, Early carbonate concretion: One example of concretionary carbonate was observed in well 10 in the South Marsh Island area (PP-142, depth=2213 m). Heavy carbon values for both siderite (in the concretion) and calcite and siderite (in the host rock) suggest that precipitation was associated with reduction of organic matter, as described by Curtis (1978). Oxygen values for calcite in this sample would permit precipitation in either brackish water at about 20^{0} C or precipitation from slightly enriched water at slightly elevated temperature. The fractionation equation for siderite is disputed (cf. Fritz and others, 1971 and Hangari and others, 1980), so conditions of siderite precipitation cannot be similarly evaluated. Because major reduction of organic matter occurs primarily within the first few hundreds of meters of burial (Curtis, 1978), a fairly early age for these nodules seems likely. This is supported by the fact that carbonate comprises over 65 percent of the rock (based on point-count data) and, lacking evidence of extensive grain replacement, must have been introduced prior to extensive compaction. Strontium isotopic evidence is consistent with siderite precipitation essentially concomitant with deposition in the late Pliocene ($^{87}Sr/^{86}Sr$=0.7089) followed by calcite precipitation somewhat

later in the burial history ($^{87}Sr/^{86}Sr$=0.7084) from water influenced by dissolution of older Tertiary nannofossils.

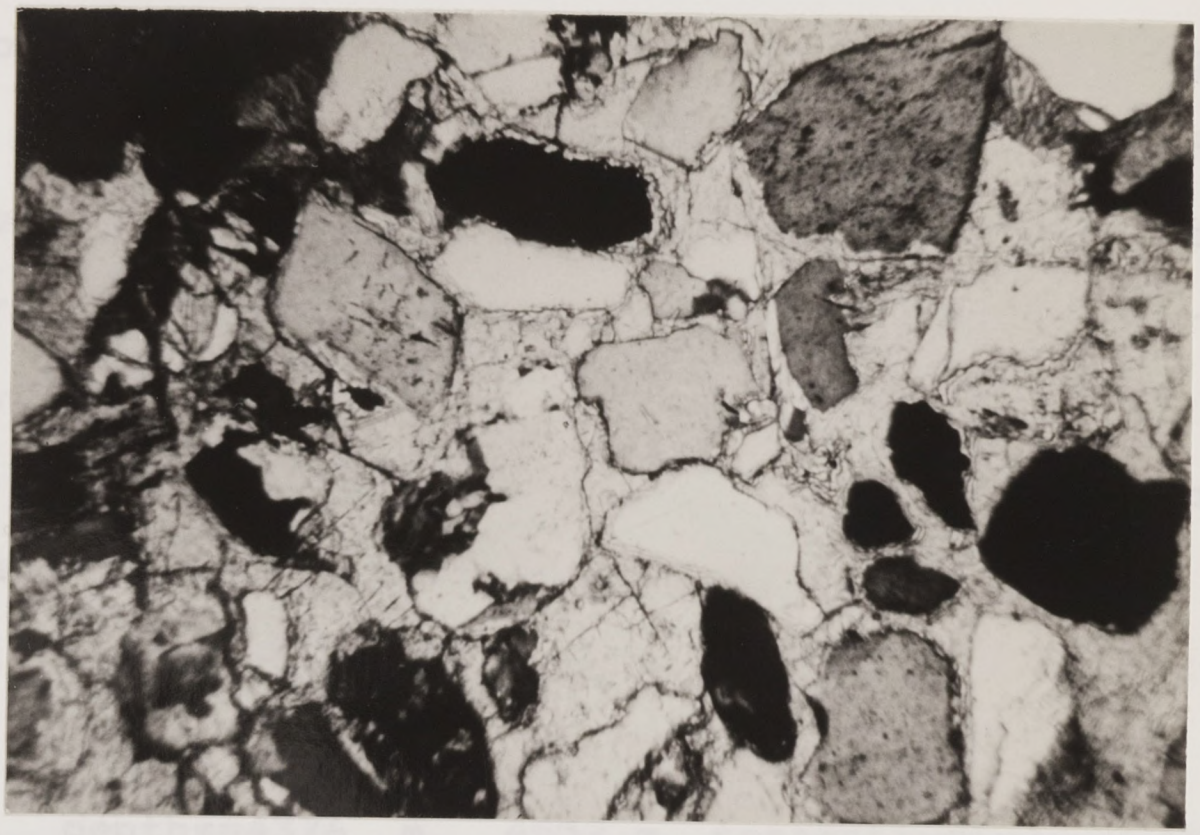
Petrographically, calcite in the host rock is undulose and nearly fibrous. Siderite in the host rock occurs as 15 to 30 um, brown rhombs. Detrital carbonate grains in the rock are coated by uniform rims of siderite (Figure 22), slight evidence that siderite precipitation indeed preceded formation of calcite. Crystal size of siderite within the nodule is variable, ranging from around 4 to 10 um. It is possible that siderite, especially where fine-grained, envelopes detrital clay, though this is difficult to evaluate.

Example 3, Nondescript calcite: Approximately 37 percent of sample PP-44 (well 3, Eugene Island area, depth <2616 m) is composed of calcite cement (Figure 23 B). Despite this high percentage of cementing material, this sample remains essentially unconsolidated. Isotopically, this sample is quite light with respect to both oxygen (-8.4 o/oo) and carbon (-12.9 o/oo). In water of $\delta^{18}O$=0 this would require a temperature of nearly 60^{0} C, which is within the realm of possibility for this sample, though lighter water at a lower temperature is also possible. The light carbon also suggests influence of organic-derived carbon in this



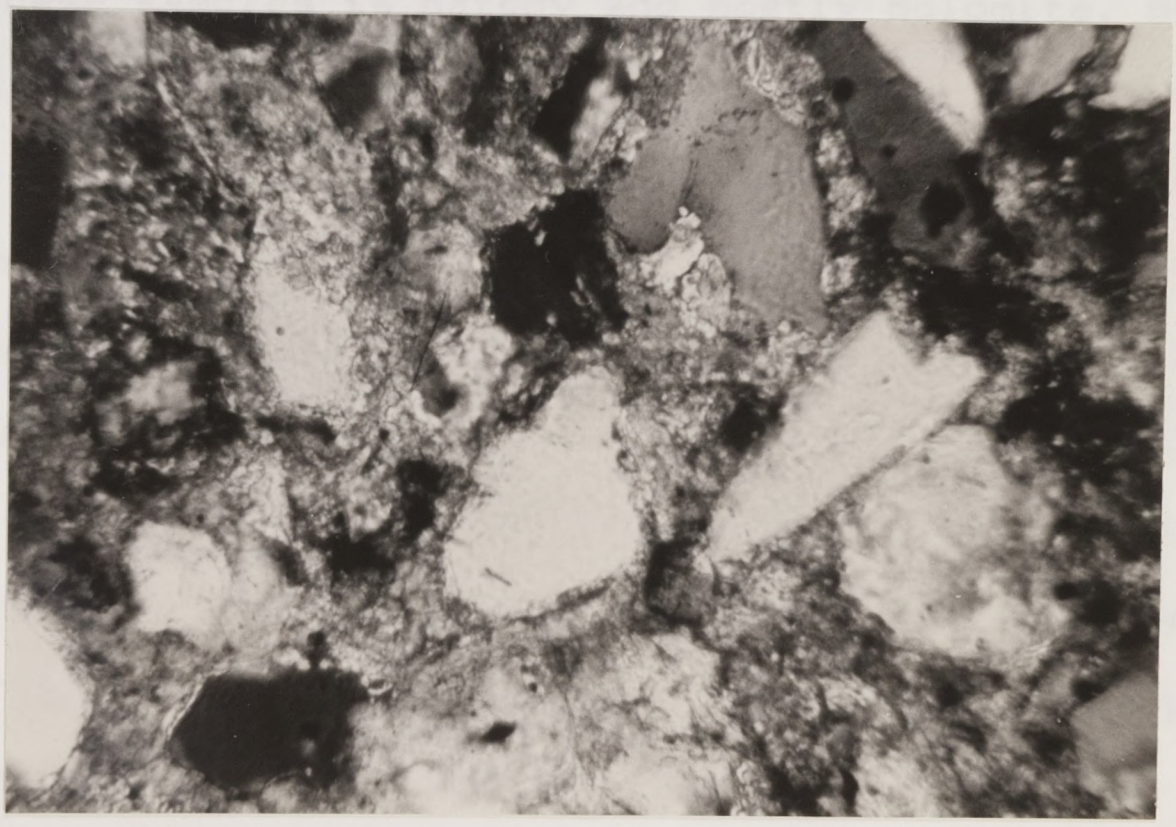
0.05 mm

Figure 22. Siderite-calcite concretion (PP-142, depth=2213 m). Plane polarized light.



A

0.2 mm



B

0.2 mm

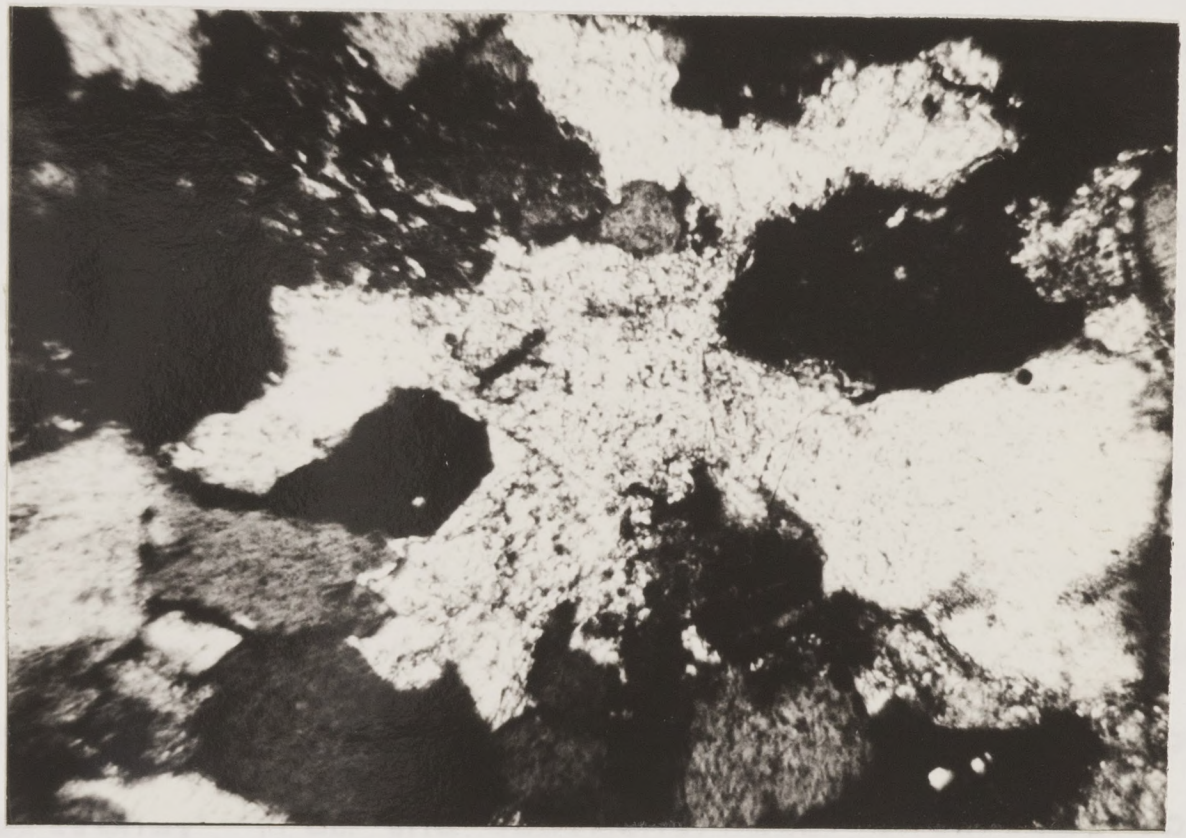
Figure 23. A. Poikilotopic calcite (PP-172, depth=2971 m). Crossed polars. B. Non-descript calcite (PP-44, depth < 2616 m). Crossed polars.

cement.

Example 4, Poikilotopic calcite: Sample PP-172 (well 11, South Marsh Island area, depth=9747) contains approximately 35 percent poikilotopic calcite (Figure 23 A). This calcite appears to be strictly a pore-filling phase; grains are not replaced. The amount of this material recovered was too small for isotopic or other analysis.

Example 5, Patchy dolomite: In some of the deepest cuttings examined in this study, highly lithified sandstone fragments contain quartz, feldspar, and carbonate cements (PP-159, P-157, South Marsh Island area, depths=4376 m, 4552 m, respectively). Carbonate in these samples is patchy and mostly dolomite. Staining shows this carbonate to be ferroan. The distribution of this dolomite suggests that carbonate precipitation occurred after significant compaction and may in substantial part represent replacement of silicates (Figure 24). Stable isotope data ($\delta^{18}O$=-6.8, $\delta^{13}C$=-4.7) also suggest that this carbonate formed at somewhat elevated temperature, as high as 80^{0} C if the water were +1. This is still somewhat lower than the present temperature (~110^{0} C) for this sample.

Carbonate minerals are the most abundant



0.1 mm

Figure 24. Late dolomite cement and/or grain-replacement (PP-159, depth=4376 m). Crossed polars.

cementing agents in the Plio-Pleistocene. This resembles the general nature of cementation in older Tertiary rocks of the Texas Gulf coast (e.g. Loucks and other, 1983; Fisher, 1983). The overall abundance of carbonate cementation, however, is much less than the averages of 5.3 percent in Frio and 3.3 percent in Wilcox sandstones (Appendix 2)(Land and Fisher, in prep.). Furthermore, siderite and dolomite are common in Plio-Pleistocene sediments while rarely found in older Tertiary units.

In general, carbonate cementation observed in this study is a highly localized phenomenon. No unified model for the origin of these carbonates can be applied; the major occurrences appear to represent a number of unique situations, encompassing a wide range of burial conditions, from the sediment/water interface to present conditions at maximum burial. Also in contrast to older Tertiary units is the observation that siderite is common in the Plio-Pleistocene.

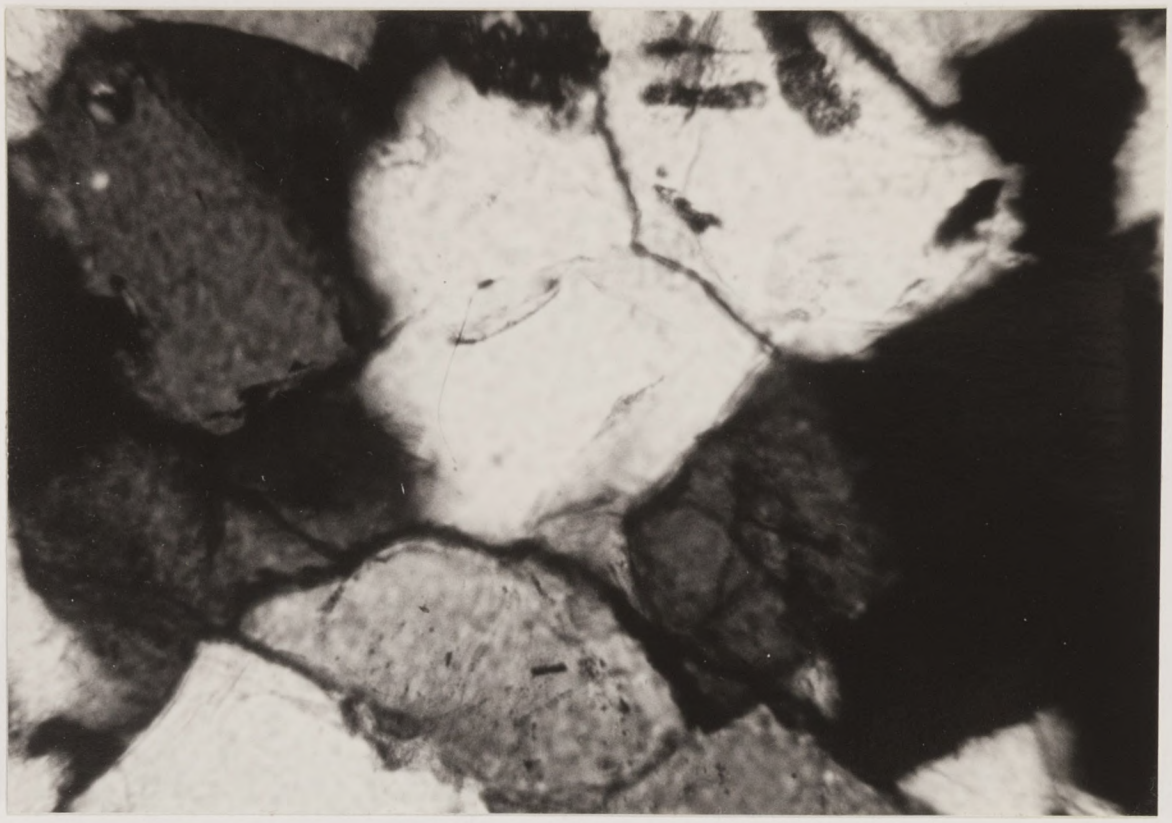
Silicate cements

Silicate cement observed in Plio-Pleistocene sediments includes quartz, potassium feldspar, albite, and kaolinite. Quartz overgrowths transported from older source rocks occur at all depths, as previously mentioned. Quartz precipitated in situ , however, is

only observed in sediments deeper than about 4 km. In contrast to transported quartz overgrowths, quartz cement is observed more or less uniformly throughout individual rock samples and is accompanied by a substantial increase in the degree of lithification. The maximum percentage of quartz cement observed is 4.4 percent (PP-159, depth=4376 m, Figure 25 A) and most samples with quartz cement contain less than one percent. Quartz cement was observed most readily with the SEM (Figure 25 B).

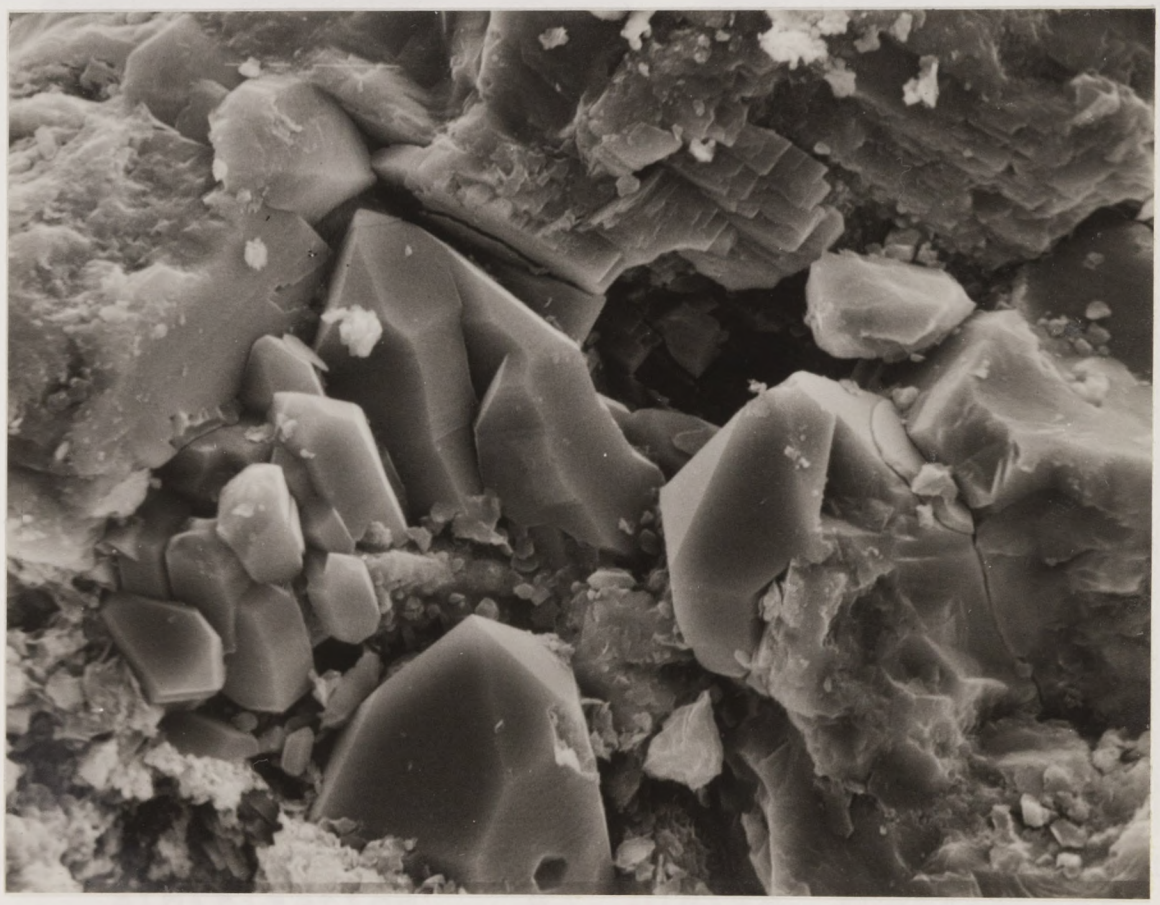
Overgrowths of both potassium feldspar and albite were also observed. Overgrowths (Fig. 8A) on detrital potassium feldspars are common, but very small, and were only observed with the SEM. Occurrence of potassium feldspar overgrowths is restricted to samples deeper than about 2000 m.

Albite overgrowths were found only on plagioclase grains and were seen both in thin section and with the SEM (Figure 26, a and c). Their occurrence is restricted to samples deeper than 3600 m. Formation of euhedral textures on detrital plagioclase grains has been suggested as evidence for a dissolution/repreciptation mechanism for albitization in sandstones (Boles, 1982). Euhedral textures characterize albitized plagioclase in the Oligocene Frio



A

0.05 mm



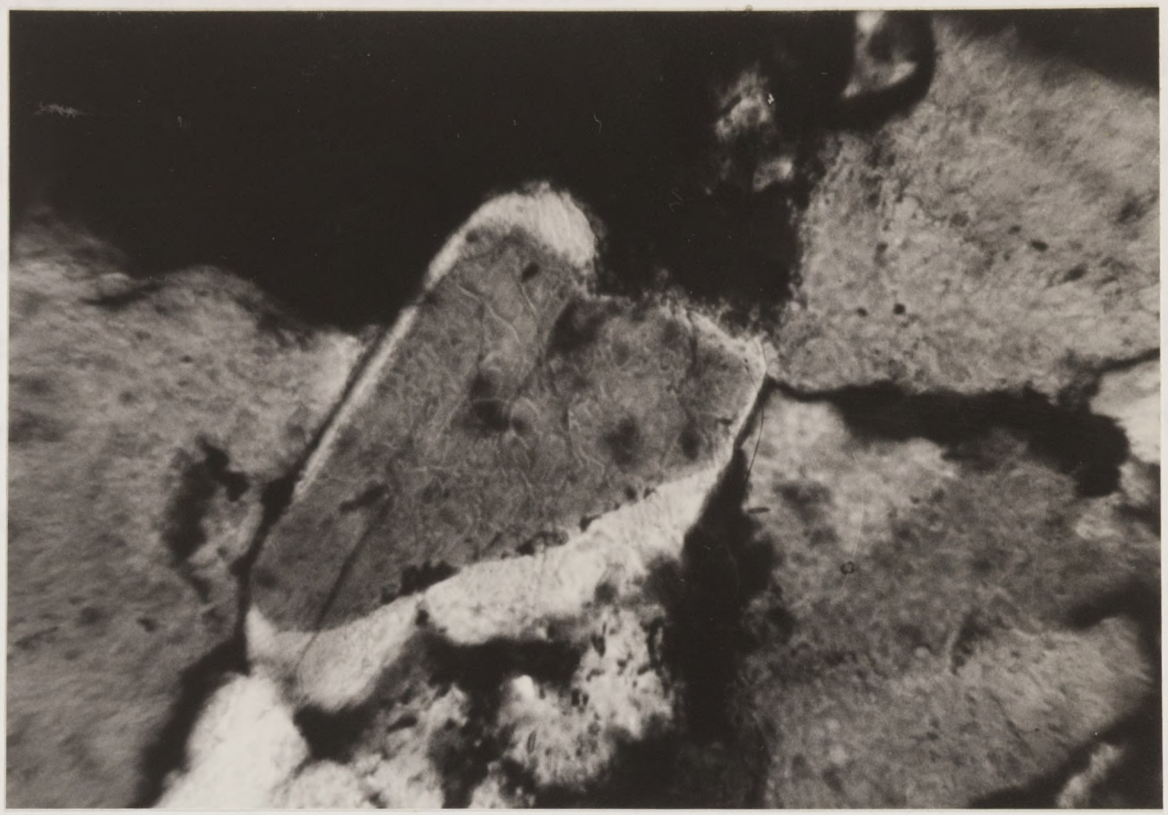
B

20 um

Figure 25. Quartz cementation. A. Extensive quartz cementation seen only in two of the deepest samples (PP-157, depth=4552 m). B. SEM view (PP-157, depth=4552 m). Crossed polars.

Figure 26. Plagioclase overgrowths associated with albitization. A. Albite overgrowths seen as clear rim on highly vacuolized and leached albitized plagioclase, Frio Formation, Brazoria County, Texas (depth >4500 m). Plane polarized light. B. Typical appearance of albitized plagioclase as seen in the SEM, Frio Formation, Brazoria County, Texas (depth >4500 m).

C



0.05 mm

D

10 um

Figure 26 continued. C. Albite overgrowth in Plio-Pleistocene sandstone (PP-157, depth=4552 m). Crossed polars. D. Aggregate of euhedral albite crystals of albitized grain in Plio-Pleistocene sample (PP-497, depth=4518 m) SEM view.

Formation and overgrowths on these grains clearly post-date partial dissolution of these grains (Figure 26, **b** and **d**). Similar appearance of overgrowths on Plio-Pleistocene plagioclase and their coincidence with the depth of incipient albitization suggests that their development in these sediments, too, is closely tied to the mechanism by which the detrital assemblage is altered.

Kaolinite cement (Figure 27) was observed most easily with the SEM. It occurs mostly in samples below about 3600 m, though not every sample below this depth contains kaolinite. Sample PP-275 (depth=4415 m) with 1.8 percent contained the most kaolinite observed. Most samples with noticeable amounts of kaolinite contain less than 0.5 percent. Kaolinite is a pore-filling phase and no relationship was noted between the presence of this cement and altered grains.

Composition of Mudstones

Primary detrital components and their alteration

Clays

In samples shallower than about 3600 m the clay mineral assemblage observed in this study is essentially

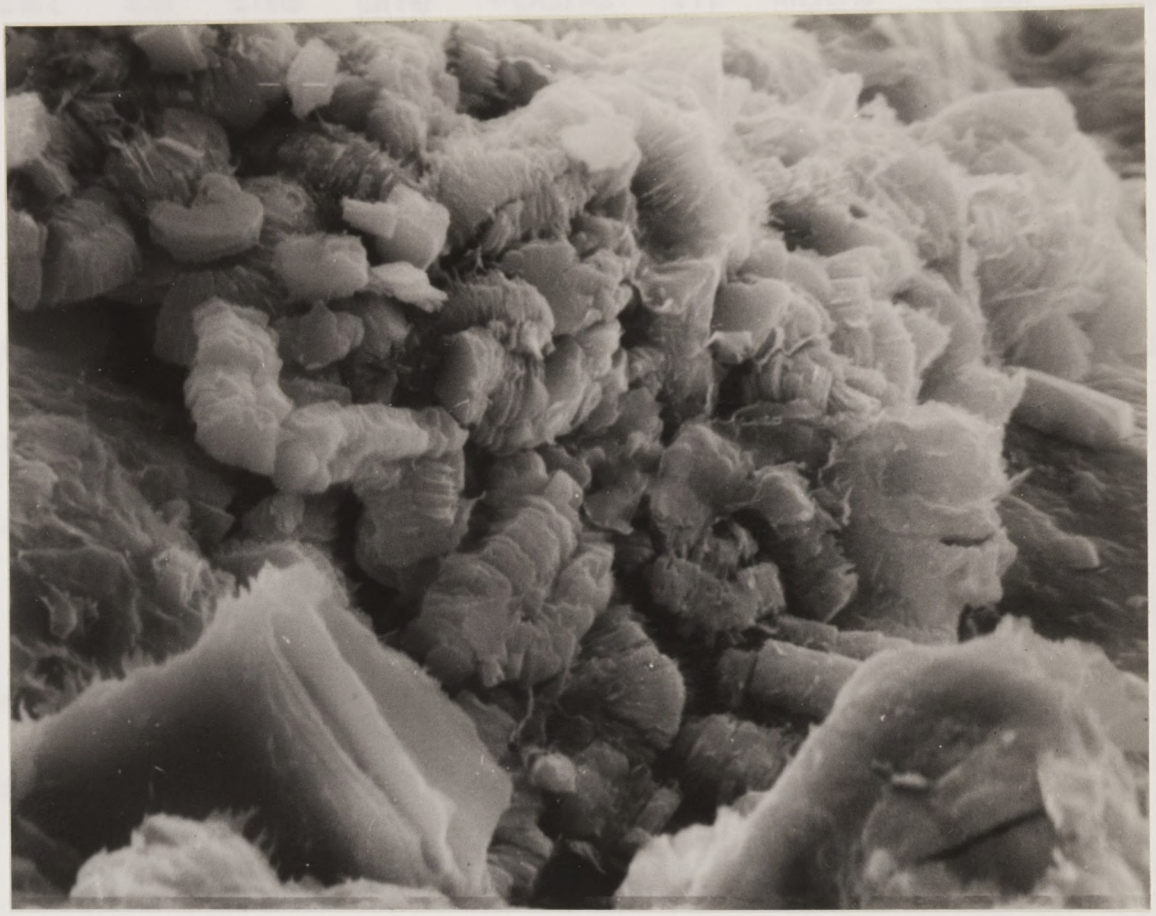


Figure 27. Kaolinite cementation. A. SEM view of vermicular kaolinite (PP-158, depth=4288 m) B. Pore-filling kaolinite seen in thin section. Cross-polars. (PP-275, depth=4415 m).

identical to the one found in muds from the lower Mississippi River and on the modern shelf (cf. Johns and Grim, 1958; Pinsak and Murray, 1960; Taggart and Kaiser, 1960; Scafe and Kunzel, 1971; Potter and others, 1975). Estimates from these papers and this study suggest that the assemblage is dominated by smectite (50%) and detrital illite (30%) with a lesser amount of kaolinite (15%) and a very small percentage of chlorite (5%) (Figures 28 and 29) and shows scant variability either geographically or with depth.

Beginning in the depth range of 3600 to 3900 m, smectite in some samples has undergone transformation to a mixed-layer smectite/illite which progressively acquires an increasing proportion of illite layers with increasing temperature (Figures 29 and 30). This reaction is accompanied by an increase in potassium (Figure 31), presumably garnered from potassium-bearing phases in the coarser fraction of the mudstone or from sandstones. Other depth-related trends in composition are decrease in total iron and increase in Al_2O_3 (Figure 32), both also logically interpreted as the results of changing proportions of illite and smectite layers.

Quantifying the extent of this reaction by the technique of Reynolds and Hower (1970) is difficult because of the large quantity of detrital illite that is

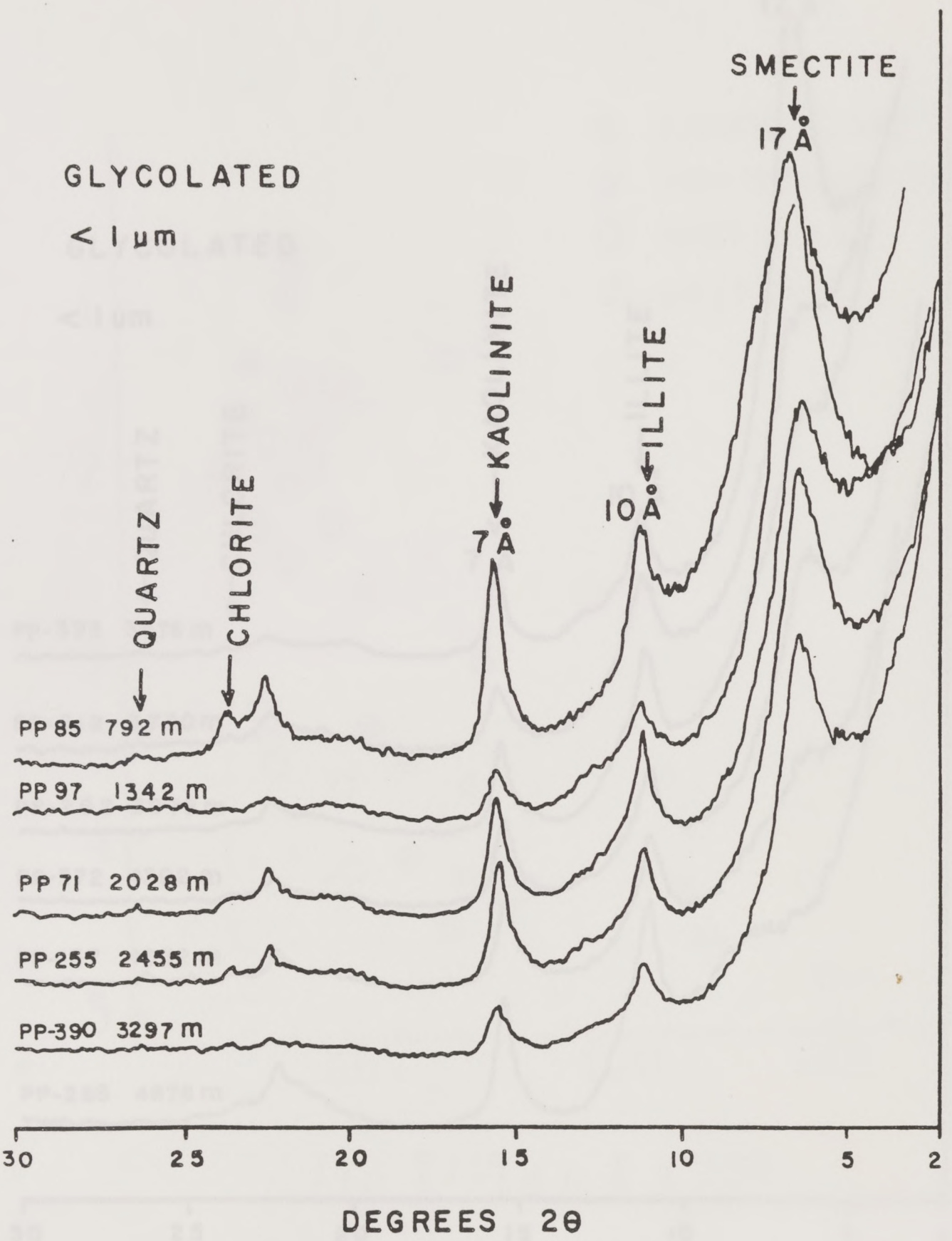


Figure 28. Diffraction patterns of unaltered clay mineral assemblages.

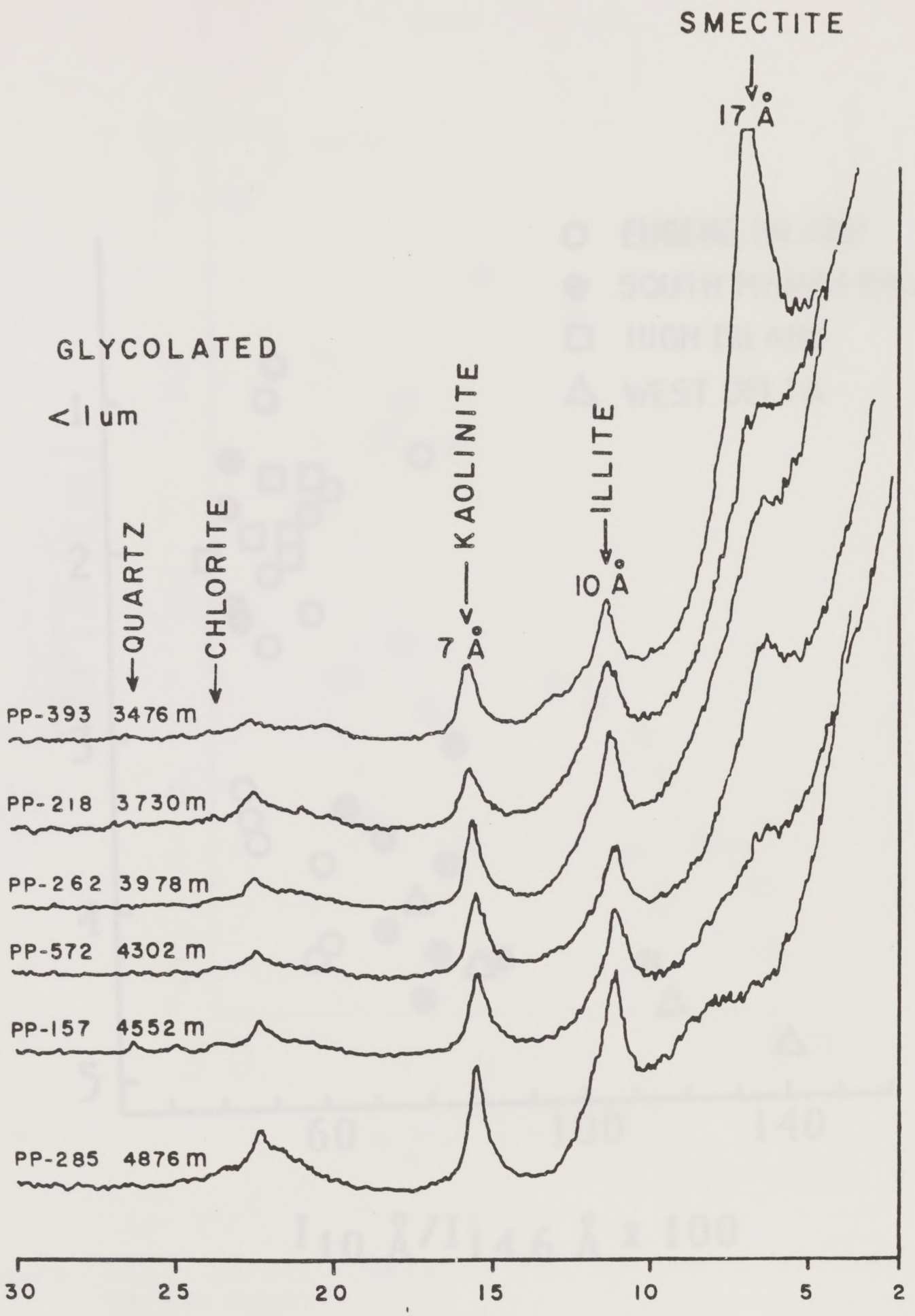


Figure 29. Diffraction patterns of deep clay samples. The 17 Å smectite peak progressively diminishes with depth and, in the deepest sample, has shifted slightly toward higher angles.

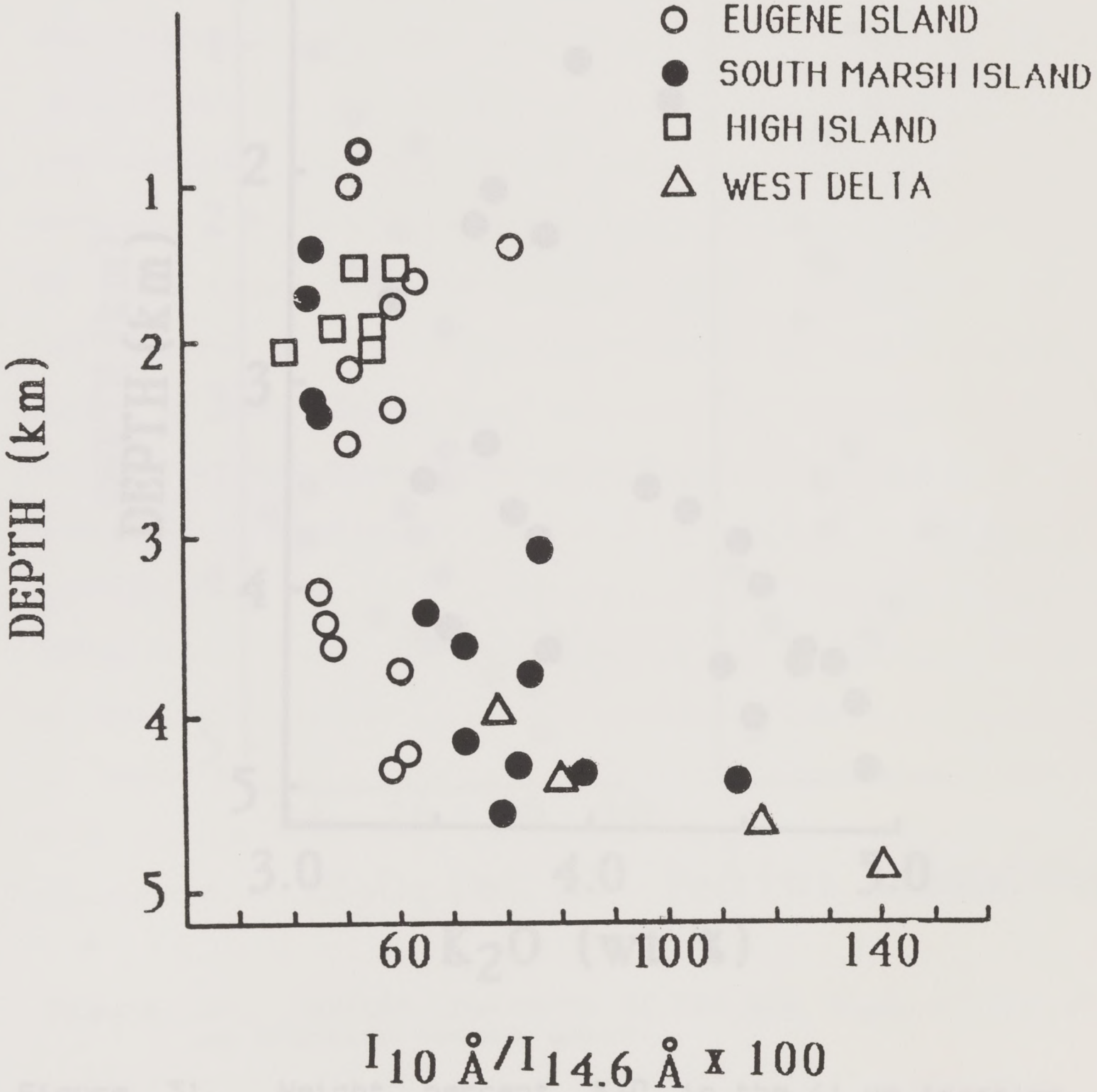


Figure 30. Ratio of 10 Å (illite) and 14 Å (smectite) peak heights versus depth.

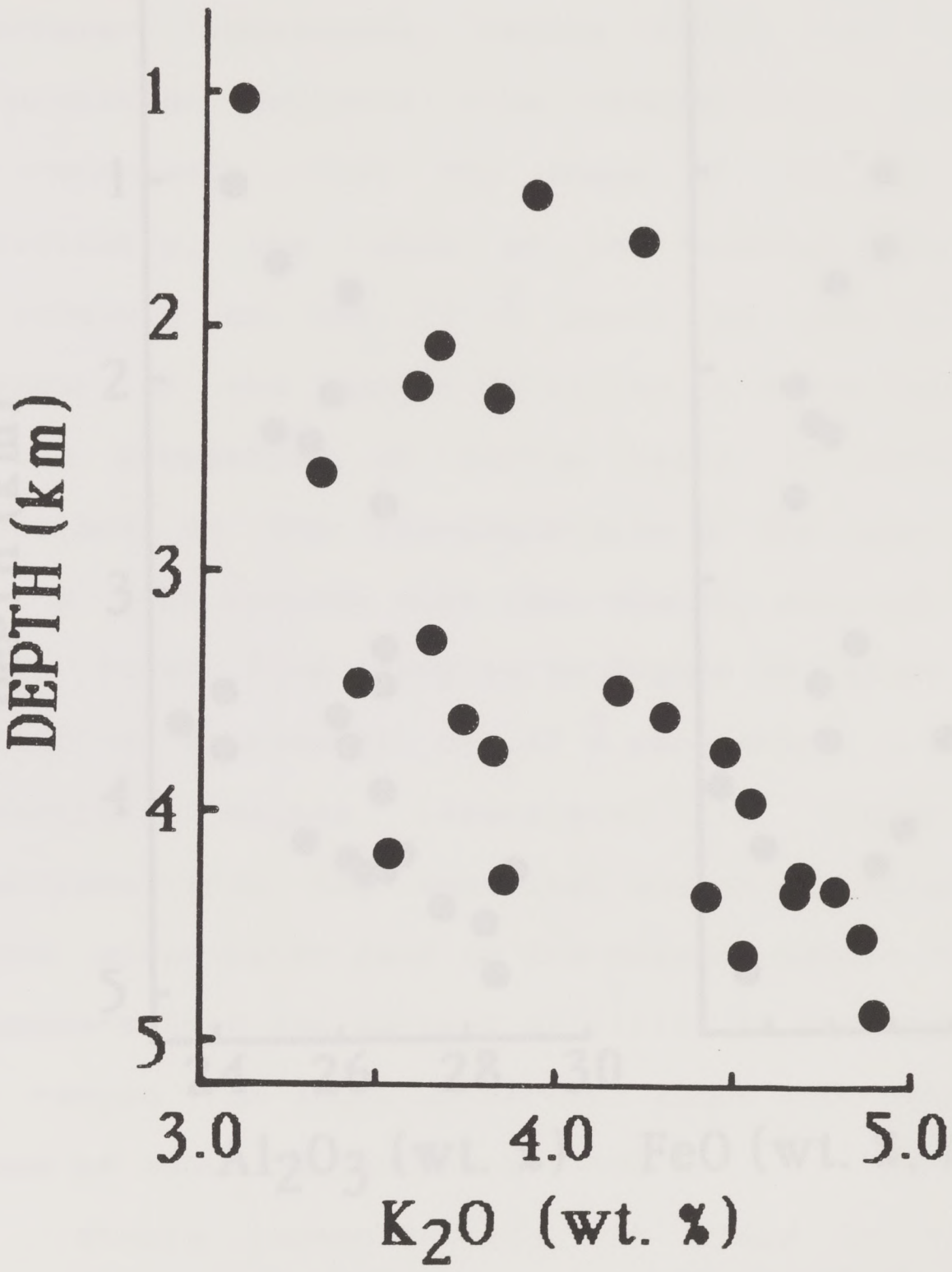


Figure 31. Weight percent K_2O in the <1 um fraction versus depth.

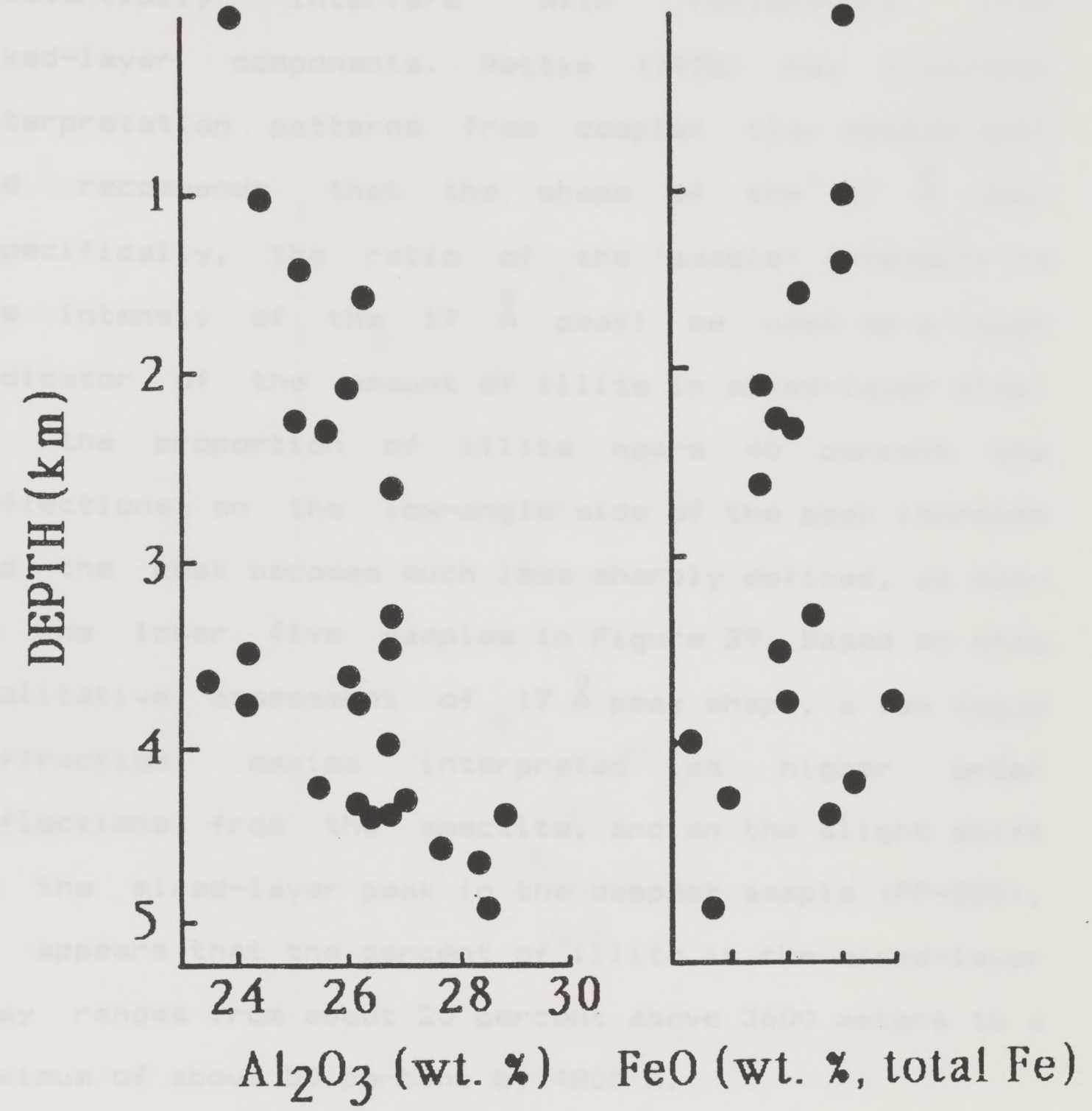


Figure 32. Weight percents of FeO and Al_2O_3 in the <1 um fraction versus depth.

present. X-ray diffraction peaks for this illite substantially interfere with reflections from mixed-layer components. Rettke (1976) has discussed interpretation patterns from complex clay assemblages and recommends that the shape of the 17 Å peak (specifically, the ratio of the "saddle" intensity to the intensiy of the 17 Å peak) be used as a rough indicator of the amount of illite in mixed-layer clay. As the proportion of illite nears 40 percent the reflections on the low-angle side of the peak increase and the peak becomes much less sharply defined, as seen in the lower five samples in Figure 29. Based on this qualitative assessment of 17 Å peak shape, a few vague diffraction maxima interpreted as higher order reflections from the smectite, and on the slight shift of the mixed-layer peak in the deepest sample (PP-280), it appears that the percent of illite in the mixed-layer clay ranges from about 20 percent above 3600 meters to a maximum of about 50 percent by 4800 m.

Stable isotopic data for clays in the <1 um fraction (Figure 33) do not really add to our understanding of the conditions of clay reactions. Clays above and below 3600 m do not differ significantly in $\delta^{18}O$. The oxygen isotopic composition of samples at all depths can probably be best explained in terms of

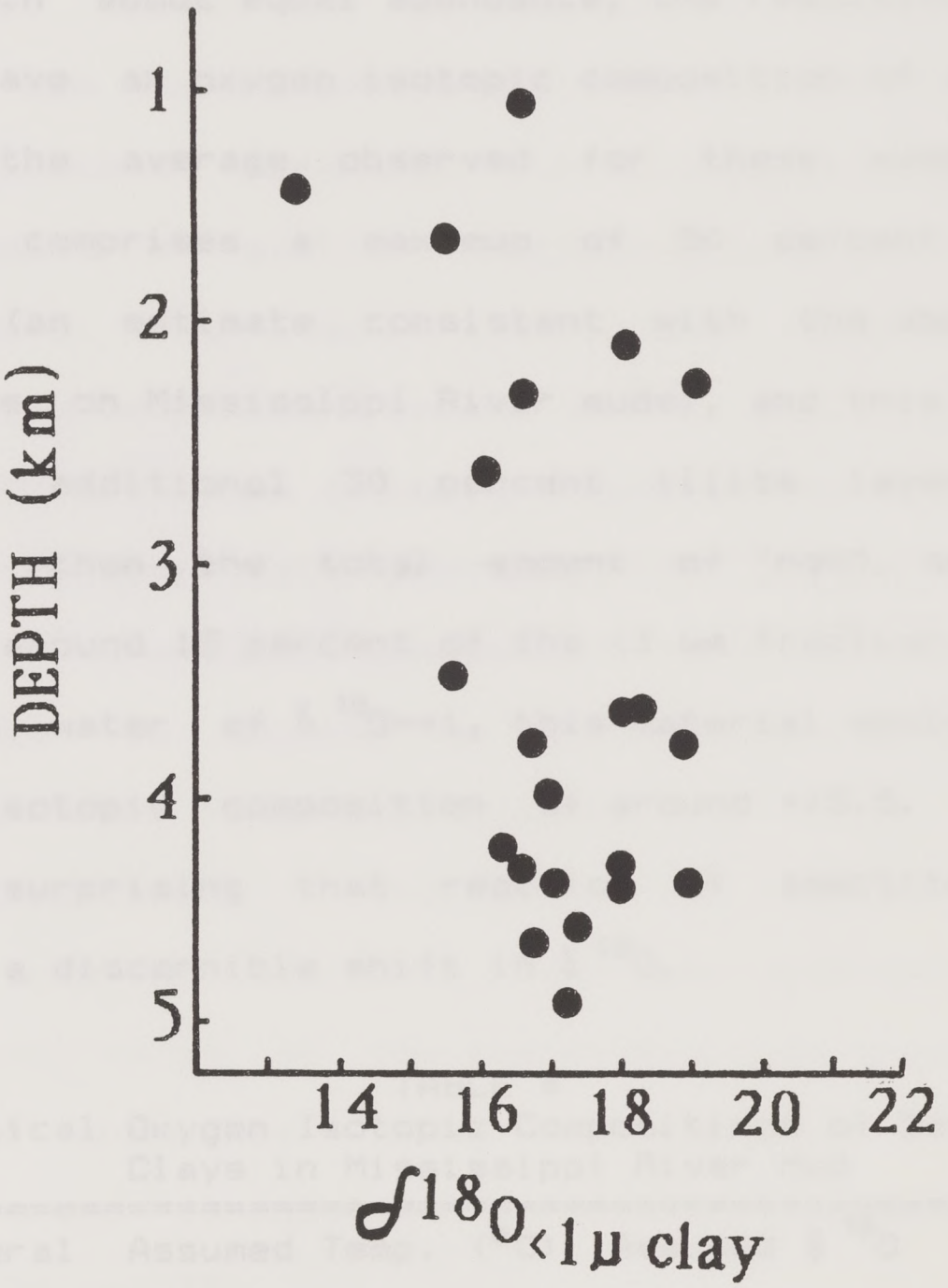


Figure 33. $\delta^{18}O$ (SMOW) of the <1 um fraction.

the isotopic compositions of the detrital clays (Table 6). If "light" clays from metamorphic source rocks (detrital illite and chlorite) and "heavy" clays formed by weathering reactions (smectite and kaolinite) are present in about equal abundance, the resulting mixture should have an oxygen isotopic composition of around 17 permil, the average observed for these samples. If smectite comprises a maximum of 50 percent of these samples (an estimate consistent with the above-cited references on Mississippi River muds), and this clay has added an additional 30 percent illite layers during reaction, then the total amount of "new", authigenic clay is around 15 percent of the <1 um fraction. At 100^{0} C, in a water of $\delta^{18}O$=+1, this material would have an oxygen isotopic composition of around +15.5. Thus, it is not surprising that reaction of smectite has not produced a discernible shift in $\delta^{18}O$.

TABLE 6

Theoretical Oxygen Isotopic Compositions of Detrital Clays in Mississippi River Mud

Mineral	Assumed Temp. (^{0}C)	Assumed $\delta^{18}O$	$\delta^{18}O$
Illite	200	+7	+13.1
Chlorite	250	+7	+ 8.0
Smectite	15	-6	+21.5
Kaolinite	15	-6	+21.5

Temperature at the beginning of the smectite to illite transition, 90^{0} to 100^{0} C, is the same or even slightly less than temperatures observed near the top of the clay transition in older Tertiary units of the Texas Gulf coast (Hower and others, 1976; Freed, 1981). However, the maximum amount of illite layers observed in Plio-Pleistocene mudstones corresponds to about the midway point of similar reactions observed in these older rocks, which acquire a maximum of about 80 percent illite by a temperature around 140^{0} C. Bruce (1984) also reports this limited extent of reaction for clays in the offshore Pliocene and ascribes the lesser degree of reaction to a smectite composition that is less reactive (beidellite as opposed to montmorillonite in older sediments of the Texas Gulf coast).

Progressive increase in chlorite and disappearance of kaolinite have been observed over the same depth range as the smectite/illite transition in Oligocene mudstones (Hower and others, 1976), but neither trend is apparent in Plio-Pleistocene mudstones (Figures 34 and 35).

Coarse fraction components

Coarse fractions (>10 um) of mudstones were examined with the SEM and with the petrographic

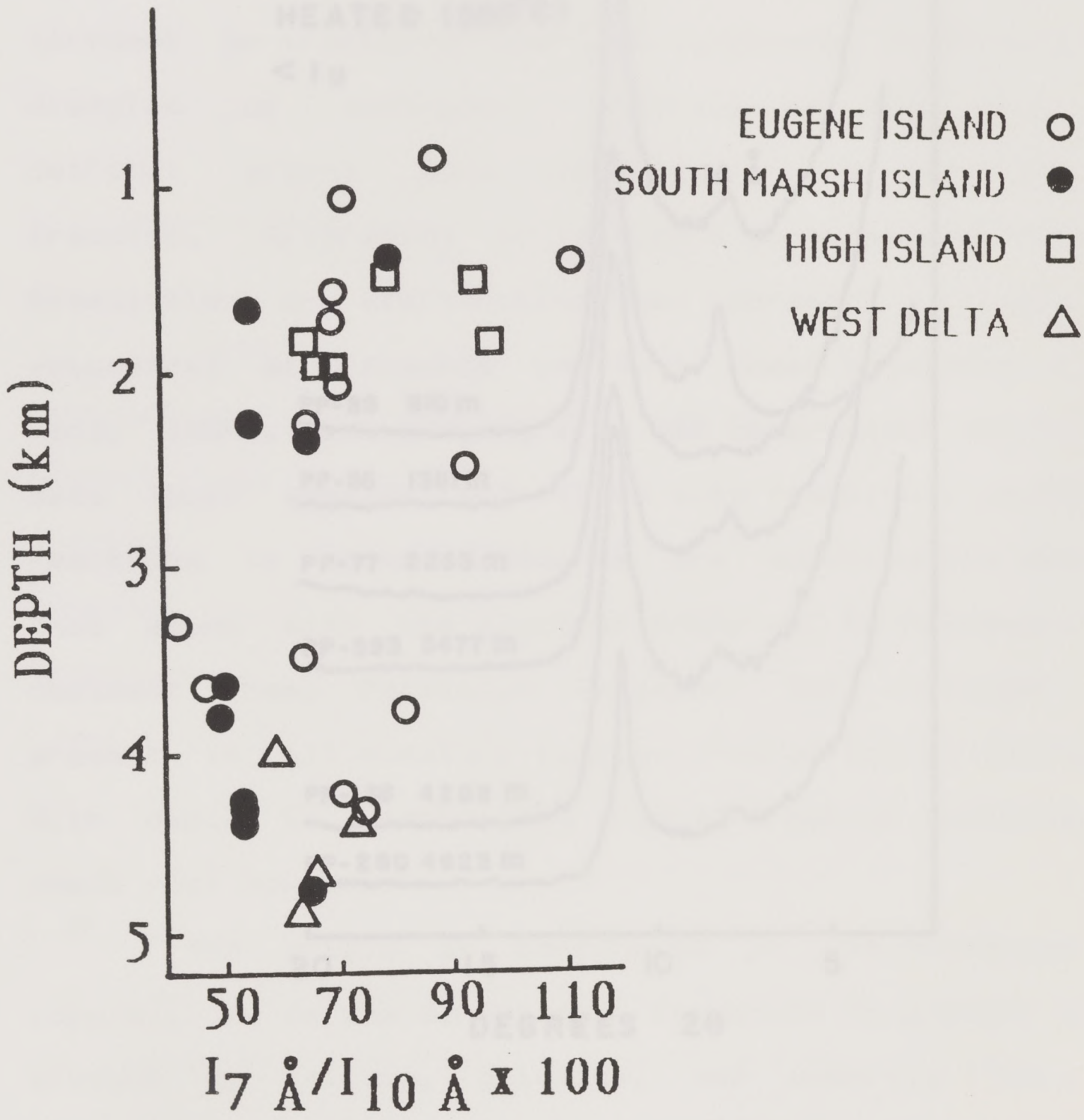


Figure 34. Ratio of 7 Å (kaolinite) and 10 Å (illite) peak heights versus depth.

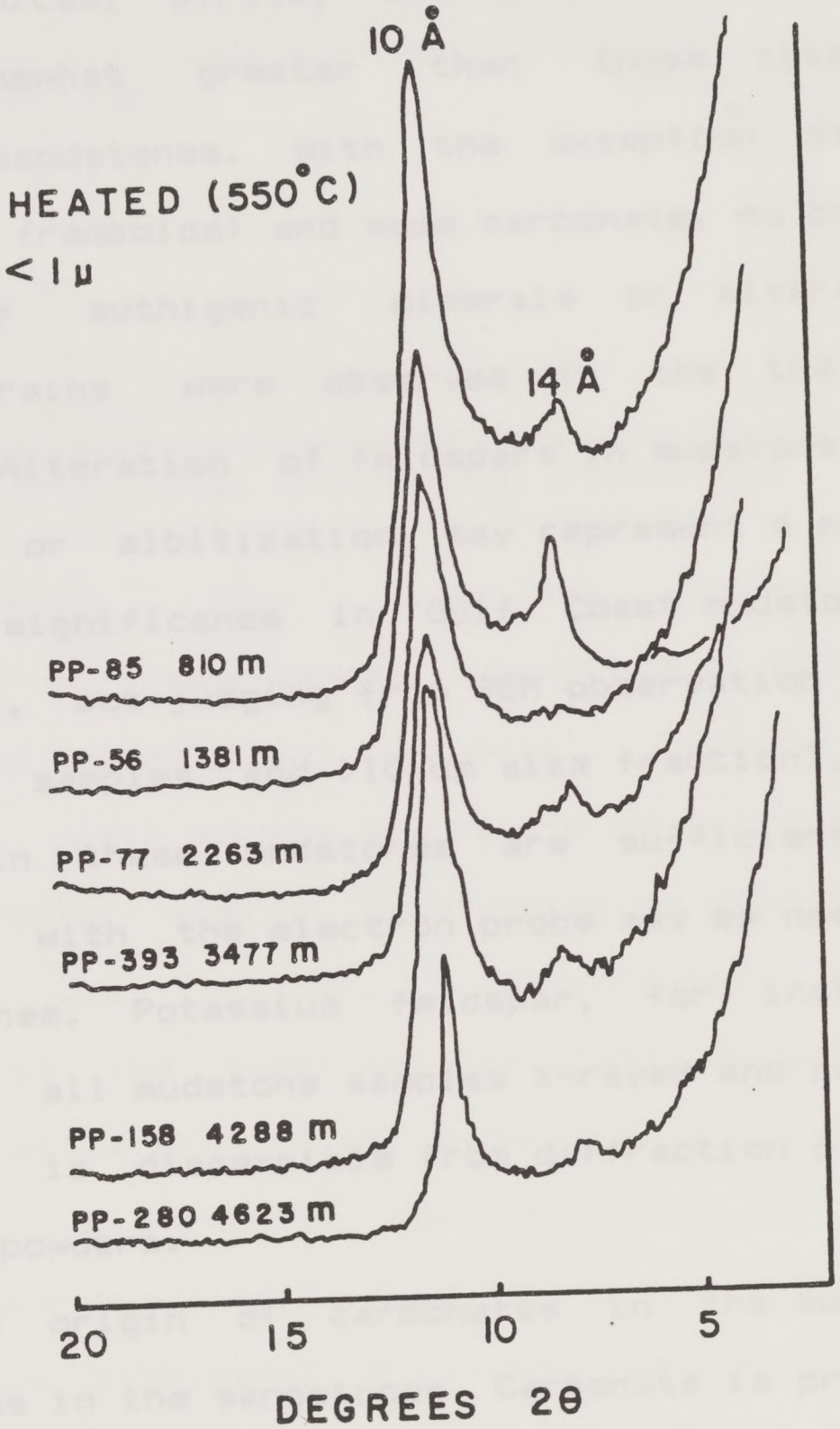


Figure 35. Diffraction patterns of representative heated (550° C) samples from different depths.

microscope, although no point counts were attemped. Mostly, this material consists of quartz, with feldspars, micas, pyrite, and carbonate all present in amounts somewhat greater than those observed in associated sandstones. With the exception of pyrite (present as framboids) and some carbonate, no convincing examples of authigenic minerals or alteration of detrital grains were observed in the the >10 um fraction. Alteration of feldspars in mudstones through dissolution or albitization may represent a process of volumetric significance in Gulf Coast mudstones (e.g. Land, 1984), but judging from SEM observation and X-ray data (bulk samples and >10 um size fraction), any such reactions in these mudstones are sufficiently subtle that study with the electron probe may be necessary to document them. Potassium feldspar, for instance, is present in all mudstone samples X-rayed and no decrease with depth is discernible from diffraction patterns of whole rock powders.

The origin of carbonates in the mudstones is complex, as in the sandstones. Carbonate is present as a mixture of calcite, dolomite, and siderite (Table 7) which represents a more complex assemblage than has been reported for older Tertiary mudstones in Texas (Hower and others, 1976, Freed, 1981; Milliken and Land, 1982;

Fisher, 1982).

TABLE 7

Abundance of Various Calcite-Dolomite-Siderite Combinations in Plio-Pleistocene Mudstones (167 samples)

Assemblage*	Percentage
C + D	17.4
C + D + S	14.4
S	13.2
D	10.8
S + D	7.8
C + S	6.0
C	4.8
No Carbonate	25.7

* C=calcite, D=dolomite, S=siderite

Clearly, the carbonates represent some combination of terrigenous detritus, allochemical debris, and authigenic material. Carbonate grains observed in grain mounts of the >10 um fraction closely resemble similar grains interpreted as detrital in the sandstones (p. 17-18). Allochemical components identified in the mudstones are mostly pelagic forams, but also include bits of molluscan debris and echinoderm parts. No signifcant concentrations of coccolith material were identified. Siderite in the mudstones must represent an authigenic component because this mineral has never been reported in Mississippi River sediment or in surface sediments of the modern shelf. Some portion of calcite and dolomite may also be authigenic, but this is difficult to evaluate in light of the known detrital

and allochemical components. Table 8 summarizes stable isotopic data for carbonates in mudstones. Neither calcite nor dolomite show any discernible isotopic trend over the depths examined. In general, carbon values for calcites are somewhat heavier than those reported for calcites in older Tertiary mudstones in Texas (Yeh and Savin, 1977; Land and Milliken, unpublished data) and oxygen values are, surprisingly, lighter. Lighter oxygen may result from the detrital component or from the effect of early brackish water diagenesis.

TABLE 8

Isotopic Composition of Carbonates in Mudstones

No.	Depth (m)	Mineral	$\delta^{18}O$	$\delta^{13}C$	% of rock
*PP-370	608	calcite	-7.5	-2.7	3.2*
PP-126	817	calcite	-6.9	-0.8	10.9
*PP-398	905	calcite	-10.0	-1.9	2.4*
*PP-399	1191	calcite	-6.9	-1.9	6.0*
*PP-373	1345	calcite	-7.9	-2.5	0.8*
PP-331	2310	calcite	-7.5	-4.1	5.7
PP-153	2311	calcite	+1.0	+0.4	3.6
*PP-382	2346	calcite	-3.3	-0.3	5.4*
PP-235	2629	calcite	-1.2	-0.7	5.5
*PP=388	3030	calcite	-7.2	-1.5	5.4*
*PP-404	3588	calcite	-4.9	-2.0	0.6*
*PP-370	608	dolomite	-1.6	-3.2	4.0*
*PP-398	905	dolomite	-1.0	-1.2	4.6*
*PP-399	1191	dolomite	-3.2	+0.6	10.6*
*PP-373	1345	dolomite	-2.0	-3.7	3.1*
PP-1	<1444	dolomite	-3.9	-6.5	7.0
PP-187	1619	dolomite	-1.6	-1.7	2.9
PP-20	<1786	dolomite	-1.9	-1.5	3.6
PP-26	<1796	dolomite	-4.0	-0.9	4.7
PP-377	1814	dolomite	-2.9	-0.9	3.7
PP-380	2025	dolomite	+1.0	-4.6	6.2
*PP-382	2346	dolomite	-2.2	+1.5	2.1*
PP-383	2463	dolomite	-2.1	-1.5	2.0
*PP-388	3030	dolomite	-2.7	-0.2	5.8*
*PP-404	3538	dolomite	+1.6	+0.2	3.9*
PP-373	1345	siderite	-6.2	+0.4	0.4
PP-103	1415	siderite	-2.7	-2.2	1.3
PP-377	1814	siderite	-4.8	+0.6	0.6
PP-380	2025	siderite	-2.2	+0.3	1.2
PP-144	2220	siderite	-2.0	+5.3	4.4
PP-383	2463	siderite	-2.2	-1.5	0.8
PP-235	2629	siderite	+2.1	+3.3	1.2

*Isotopic data obtained through double extractions. "Calcite" gas was collected after one half hour. "Intermediate" gas was discarded after 3 hours and represents as much as an additional percent of carbonate in the sample. "Dolomite" gas was collected after one week to ten days.

IMPLICATIONS OF THE PLIO-PLEISTOCENE FOR MODELS OF GULF COAST DIAGENESIS

Diagenetic modification of terrigenous sediments in the Gulf of Mexico sedimentary basin has not been uniform. This was recognized by Land (1984b) and Land and Fisher (in prep.) in regional comparisons of diagenesis in Eocene and Oligocene units along the Texas coast. These early Tertiary formations share basic characteristics; quartz and carbonate are the main cementing agents. They differ, though, in particulars of cement chemistry as well as in the distribution of cementation (depth to lithification is deeper in the Frio). This study extends this conclusion to the youngest of the Gulf's terrigenous wedges. Figure 36 summarizes the depths of occurrence of diagenetic phases and processes observed in the Plio-Pleistocene. The general nature of cementation (quartz + carbonate) and alteration of detrital minerals (albitization, dissolution, smectite/illite transformation) matches that seen elsewhere in the Gulf. The depth to cementation and grain alteration is tremendously greater.

Temperatures at the onset of these processes however are <u>not</u> especially different from temperatures

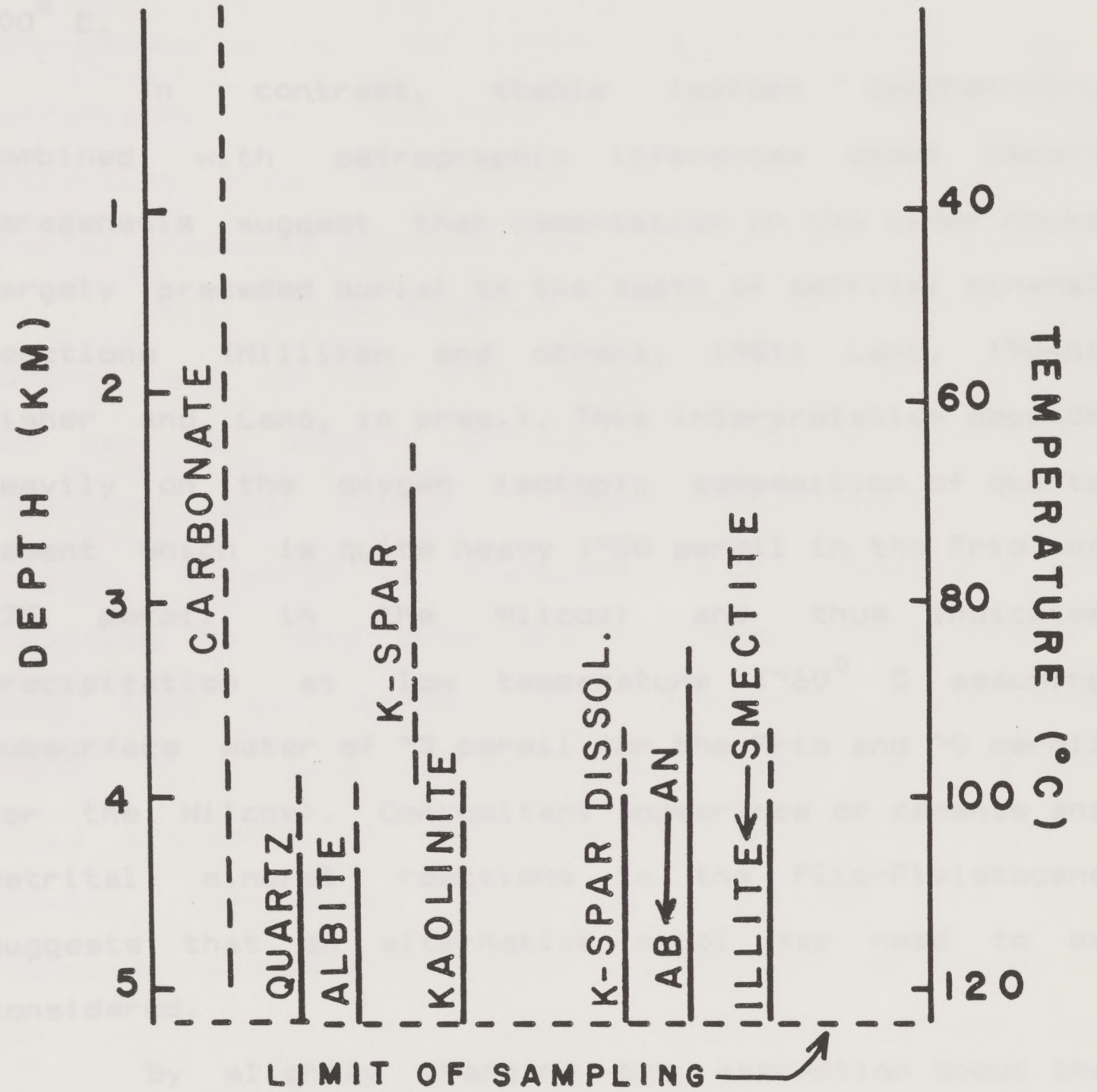


Figure 36. Depth of occurrence of diagenetic products and processes in Plio-Pleistocene sediments. Where lines are dashed the constituent is a very minor and sporadic component of the sediment; where the lines are solid the constituent may still be minor in terms of overall abundance, but it is more or less consistently present.

observed for similar degrees of diagenesis in older units. Both grain alteration and significant cementation are first observed in the realm of 90^{o} to 100^{o} C.

In contrast, stable isotope geochemistry combined with petrographic inferences about cement paragenesis suggest that cementation in the older rocks largely preceded burial to the depth of detrital mineral reactions (Milliken and others, 1981; Land, 1984b; Fisher and Land, in prep.). This interpretation depends heavily on the oxygen isotopic composition of quartz cement which is quite heavy (~30 permil in the Frio and ~25 permil in the Wilcox) and thus indicates precipitation at low temperature (~60^{o} C assuming subsurface water of ~3 permil for the Frio and ~0 permil for the Wilcox). Concomitant appearance of cements and detrital mineral reactions in the Plio-Pleistocene suggests that an alternative model may need to be considered.

By slightly changing the assumption about the isotopic composition of the fluid responsible for quartz cementation in Eocene and Oligocene sandstones, it is possible to arrive at a higher temperature of quartz precipitation (say, 100^{o} C in a water of +7 o/oo). If

this were the case, then the cementation process and detrital mineral alterations could be viewed as spatially related events in each of the Gulf Coast terrigenous units. The onset of both processes closely corresponds to lithification and all three rock properties track one another to increasing depths as thermal gradients decrease from older to younger units.

This scheme suggests, but does not require, that cements are derived from materials released by one of the detrital mineral reactions and likewise, that these reactions are rather temperature-dependent. Fluid volumes required to transport cement materials are unchanged by this model and the "water-volume problem" (e.g. Land and Dutton, 1979, Bjorlykke, 1979; Land, 1984b) still must be resolved. Indeed, long-distance fluid transport of materials from underlying rocks of varying composition could be integrated with this "temperature-dependent" model to explain variations in the particulars of cement chemistry between the different units.

SUMMARY OF CONCLUSIONS

1. Plio-Pleistocene sands beneath the Louisiana shelf are mostly lithic arkoses and feldspathic litharenites. In unaltered sands plagioclase (average anorthite content around 25 percent) is the dominant feldspar, typically comprising about 50 to 60 percent of the total feldspar. The overall primary detrital assemblage closely resembles the detritus carried by the modern Mississippi River and in terms of mineralogical stability equals or exceeds the potential reactivity of primary detrital assemblages in older Tertiary sediments of the Texas Gulf coast.

2. Detrital clay minerals found in the <1 um fraction of mudstones include smectite, illite, kaolinite, and chlorite. Again, this detrital assemblage matches closely the one reported for the modern Mississippi.

3. Dissolution of potassium feldspar in sands occurs in a zone from approximately 3700 m to 5000 m, at temperatures between about 95^{0} and 120^{0} C.

4. Incipient albitization is evident in a few of the deepest samples (below about 4600 m) and may also

have affected samples in the depth range of 3300 to 4600 m.

5. Alteration of smectite by addition of illite layers occurs below about 3500 to 4000 m but does not advance beyond about 50 percent illite in illite/smectite at 5000 m.

6. Burial diagenesis is a very effective agent for modification of detrital assemblages. More alteration of the primary detrital assemblage has occurred during 5 km of burial over 5 million years than in several thousand miles of transport down the Mississippi River system (Russel, 1937).

7. As in older Tertiary formations of the Texas Gulf coast, cementation in the Plio-Pleistocene is dominated by carbonates and quartz. In contrast to the older sandstones however, carbonate cementation is rare, highly localized, and includes a more complex assemblage of carbonates. A substantial portion of the dolomite observed in the Plio-Pleistocene is detrital. Also in contrast to older units is the greater depth (approximately 4 km) to quartz cementation. Above this depth, the

majority of samples are consolidated by clay matrix or not at all.

8. Carbonates in mudstones, which include the same complex assemblage found in sandstones, may also be somewhat less abundant in comparison to mudstones in the older Tertiary rocks of the Texas Gulf coast.

9. This study of the Plio-Pleistocene confirms that there has been an evolution of diagenesis during filling of the Gulf of Mexico. It is clear that the various wedges of Tertiary sediment are characterized by substantial differences and, to a degree, these differences vary systematically from the older to the younger units and from the basin edge to basin center.

10. A uniform model to explain this systematic variation has not yet been devised. Such a model will require better understanding of thermal history as well as more information on the composition of the older sediments onto which the Tertiary wedges were deposited.

APPENDIX 1.

Point-count data *

NUMBER	DEPTH 1	DEPTH 2	Q	F	R	NON-QFR	P/P+K	A	M	P	%F	SIZE	SRF	VRF	MRF
EUGENE ISLAND															
PP-439	1454	443	70.70	13.30	16.00	-----	0.59	0.00	-----	-----	----	1.25	72.20	5.60	22.20
PP-430	1512	461	59.00	23.40	17.60	0.73	0.61	41.40	0.00	4.10	12.60	2.75	36.40	21.20	42.40
PP-449	1867	569	72.00	15.90	12.10	0.25	0.53	43.70	0.00	2.00	8.60	2.00	43.70	31.30	25.00
PP-371	2091	637	74.40	11.80	13.70	1.09	0.64	0.00	58.00	2.50	4.50	2.25	55.60	14.80	29.60
PP-87	2616	797	74.10	14.40	11.50	0.00	0.36	0.00	29.40	19.00	6.80	2.00	65.00	15.00	20.00
PP-92	3285	1001	76.70	12.60	10.70	1.21	0.45	0.00	9.10	26.30	6.30	1.50	75.00	12.50	12.50
PP-93	3300	1006	81.40	8.20	10.40	0.22	0.53	0.00	16.10	49.40	4.10	1.50	64.70	5.90	29.40
PP-57	4662	1421	68.80	15.60	15.60	1.46	0.41	0.00	16.70	21.90	9.40	3.00	75.00	25.00	0.00
PP-59	4931	1503	64.40	20.70	14.90	0.75	0.67	0.00	18.10	25.40	11.60	2.75	40.90	22.70	36.40
PP-60	4932	1503	57.10	18.30	24.60	1.30	0.72	1.50	18.90	21.00	10.40	3.00	50.00	29.50	20.50
PP-66	5345	1629	59.70	22.50	17.80	4.00	0.76	8.80	9.40	26.50	11.50	4.00	30.00	20.00	50.00
PP-74	6681	2036	62.10	15.50	22.40	-----	0.58	0.00	-----	-----	----	3.50	52.60	10.50	36.80
PP-81	7456	2273	64.00	22.30	13.70	1.52	0.51	13.40	0.60	20.40	14.30	3.00	42.90	28.50	28.60
PP-194	6264	1909	62.90	24.40	12.70	0.23	0.36	0.00	38.90	12.50	11.80	3.50	30.40	30.50	39.10
PP-196	6283	1915	76.80	16.00	7.20	0.00	0.42	0.00	0.00	33.00	10.80	2.00	84.60	15.40	0.00
PP-40	8575	2614	63.00	19.20	17.80	0.21	0.43	0.30	27.00	13.80	10.90	4.00	81.80	4.50	13.60
PP-44	8584	2616	59.50	27.90	12.60	1.43	0.58	36.70	0.20	9.30	14.80	3.75	38.90	38.90	22.20
PP-49	8602	2622	55.80	29.10	14.10	0.43	0.67	0.00	50.60	5.00	12.90	4.00	90.90	0.00	9.10
PP-50	8605	2623	61.40	22.40	16.20	1.85	0.61	0.00	47.40	7.60	9.60	3.75	21.40	53.60	25.00
PP-54	9167	2794	64.40	18.50	17.10	2.00	0.53	0.00	49.60	10.50	----	3.75	86.70	13.30	0.00
PP-763	11467	3495	58.00	23.90	18.10	0.29	0.64	0.00	0.00	30.70	16.50	3.00	26.50	29.40	44.10
PP-762	11808	3599	55.70	25.10	19.20	0.27	0.67	0.00	1.10	39.70	14.90	3.25	15.40	59.00	25.60
PP-406	11882	3622	57.70	23.40	18.90	0.71	0.69	0.00	7.10	39.90	12.30	3.25	45.40	27.30	27.30
PP-751	12302	3750	43.50	25.90	30.60	0.12	0.59	0.00	6.20	24.40	17.60	3.50	34.30	42.80	22.90
PP-748	12354	3765	45.50	23.40	31.10	0.89	0.67	0.00	3.50	29.90	15.40	3.50	46.80	27.40	25.80
PP-747	12358	3767	48.80	30.90	20.30	1.70	0.81	0.48	25.90	21.80	15.50	3.75	24.20	39.40	36.40
PP-746	12362	3768	49.60	29.40	21.00	1.38	0.83	0.22	23.80	22.20	15.40	3.50	50.00	25.00	25.00
HIGH ISLAND															
PP-12	5840	1780	55.40	20.50	24.50	1.84	0.70	0.30	10.80	29.70	11.80	3.75	83.90	3.20	12.90
PP-13	5847	1782	49.80	26.50	23.70	3.87	0.62	0.00	6.10	49.60	10.70	4.00	40.70	22.20	37.10
PP-17	5854	1784	58.70	19.90	21.40	5.69	0.65	0.00	19.70	27.00	9.50	4.00	100.00	0.00	0.00

NUMBER	DEPTH 1	DEPTH 2	Q	F	R	NON-QFR	P/P+K	A	M	P	%F	SIZE	SRF	VRF	MRF
SOUTH MARSH ISLAND															
PP-101	4632	1412	73.90	9.90	16.20	-----	0.50	0.00	-----	-----	----	2.00	93.50	0.00	6.50
PP-138	4704	1434	75.60	14.60	9.80	-----	0.71	0.00	-----	-----	----	2.50	54.50	18.20	27.30
PP-140	4743	1446	66.50	19.60	13.80	0.61	0.48	0.00	1.50	30.40	13.40	2.25	48.10	48.20	3.70
PP-106	5176	1578	64.40	14.20	21.40	0.44	0.72	0.00	0.96	44.10	7.70	2.75	71.10	17.80	11.10
PP-109	5194	1583	66.20	21.60	11.70	-----	0.57	0.00	-----	-----	----	2.50	36.40	18.20	45.40
PP-113	5298	1615	79.30	12.80	7.80	-----	0.32	0.00	-----	-----	----	2.00	80.00	20.00	0.00
PP-148	7419	2261	62.90	20.40	16.70	-----	0.71	0.00	-----	-----	----	3.25	25.00	53.10	21.90
PP-151	7575	2309	65.00	15.90	19.10	2.65	0.74	0.00	12.40	27.20	9.40	3.00	87.50	6.20	6.30
PP-142	7261	2213	63.00	21.70	15.30	0.71	0.51	68.80	0.00	0.00	6.60	4.00	37.50	37.50	25.00
PP-152	7579	2310	72.70	16.20	11.10	-----	0.41	1.90	-----	-----	----	2.75	65.20	8.70	26.10
PP-172	9747	2971	61.10	22.20	16.70	1.46	0.52	34.70	0.87	0.00	14.00	2.75	75.00	18.70	6.30
PP-176	12140	3700	62.30	18.90	18.80	-----	0.73	0.00	-----	-----	----	3.75	50.00	31.20	18.80
PP-186	12859	3919	67.30	17.50	15.20	0.69	0.79	0.20	25.00	22.50	9.00	3.00	38.50	19.20	42.30
PP-158	14069	4288	66.40	11.40	22.30	0.63	0.60	3.50	26.30	0.00	7.90	3.00	42.10	10.50	47.40
PP-159	14357	4376	69.50	21.90	8.60	1.70	0.85	21.70	5.40	0.00	15.60	4.00	77.80	11.10	11.10
PP-157	14933	4552	71.30	15.70	13.00	0.00	0.76	24.20	5.50	0.65	11.00	3.25	5.90	52.90	41.20
WEST DELTA															
PP-788	8046	2452	74.90	12.60	12.50	-----	0.35	0.00	-----	-----	----	2.50	60.00	16.00	24.00
PP-787	8059	2456	78.00	13.60	8.40	-----	0.39	0.00	-----	-----	----	2.25	35.30	47.10	17.60
PP-783	8199	2499	78.00	14.50	7.50	-----	0.62	0.00	-----	-----	----	2.25	38.90	27.80	33.30
PP-308	10929	3331	74.90	15.50	9.60	0.52	0.48	0.00	5.80	36.20	8.90	3.25	35.30	11.80	52.90
PP-309	10934	3333	79.80	14.30	5.90	0.00	0.54	1.30	21.40	24.80	7.70	2.50	44.40	11.10	44.40
PP-247	10942	3335	58.40	27.90	13.70	3.57	0.73	0.00	42.90	3.30	14.00	3.25	28.60	66.70	4.70
PP-248	10951	3338	59.80	13.70	26.50	1.99	0.52	0.00	0.00	31.80	9.10	2.25	27.70	46.80	25.50
PP-252	10965	3342	59.70	16.50	23.90	0.59	0.78	0.00	1.50	26.50	11.80	3.50	27.30	15.10	57.60
PP-253	10969	3343	70.10	10.40	19.50	0.00	0.50	0.00	5.70	22.30	7.50	3.00	50.00	36.70	13.30
PP-474	11055	3370	65.20	23.70	11.10	0.32	0.51	1.00	15.10	17.60	15.70	3.00	38.10	14.30	47.60
PP-312	12569	3831	54.80	18.00	27.20	0.23	0.76	0.00	2.80	44.50	9.50	3.25	39.50	39.50	21.00
PP-260	12590	3837	65.80	14.40	19.80	0.50	0.59	0.00	10.40	33.80	8.00	3.75	64.50	6.50	29.00
PP-261	13046	3976	72.90	20.90	6.20	4.34	0.63	0.00	52.50	0.00	9.00	3.25	50.00	50.00	0.00
PP-273	14448	4404	66.90	17.00	16.10	2.27	0.65	0.00	53.00	0.00	7.60	3.75	50.00	31.20	18.80
PP-275	14484	4415	60.30	13.10	26.60	1.19	0.73	2.10	2.70	25.90	8.90	3.00	27.50	20.00	52.40
PP-278	14592	4448	74.30	16.50	9.20	0.00	0.68	0.00	20.20	24.70	9.00	3.00	50.00	28.60	21.40
PP-288	16556	5046	80.10	13.10	6.80	3.34	1.00	0.00	47.40	0.00	6.50	4.00	87.50	0.00	12.50

NUMBER	DEPTH 1	DEPTH 2	Q	F	R	NON-QFR	P/P+K	A	M	P	%F	SIZE	SRF	VRF	MRF
WEST CAMERON															
PP-129	2728	831	71.20	15.00	13.70	0.24	0.61	0.00	3.60	41.10	8.30	2.25	30.80	38.40	30.80
PP-130	2741	835	62.10	17.10	20.80	0.00	0.41	0.00	5.90	39.60	9.20	2.25	44.00	28.00	28.00
PP-132	2768	844	75.40	15.80	8.80	0.91	0.71	0.00	6.40	40.60	8.20	2.00	38.90	33.30	27.80
PP-118	7314	2229	54.60	22.00	23.40	-----	0.58	0.00	-----	-----	----	2.75	60.70	21.40	17.00
PP-119	7323	2232	72.20	14.40	13.40	-----	0.67	0.00	-----	-----	----	2.50	58.80	0.00	41.20
SHIP SHOAL															
PP-8268	8268	2520	78.00	9.90	12.00	-----	0.26	0.00	-----	-----	----	2.00	84.60	5.80	9.60
PP-304	8899	2712	59.00	24.30	16.70	1.35	0.41	1.30	24.10	32.60	9.80	3.00	28.60	67.90	3.50
PP-770	8981	2737	56.40	20.40	23.20	0.84	0.53	0.00	40.30	12.10	9.60	3.50	38.10	42.90	19.00
PP-771	8983	2738	60.10	15.90	24.00	----	0.46	0.00	-----	-----	----	3.25	81.50	18.50	0.00
PP-772	8984	2738	60.10	16.70	23.20	1.47	0.68	0.99	2.20	39.80	9.30	3.50	40.90	31.80	27.30
PP-773	8999	2743	50.20	28.40	21.40	1.72	0.65	0.00	25.00	21.40	15.90	3.75	50.00	27.50	22.50
PP-776	9014	2747	57.50	21.50	21.00	1.40	0.67	0.00	0.56	36.50	13.20	3.75	38.20	35.30	26.50
MISS R.	0.00	0	64.80	16.70	18.50	-----	0.60	0.00	-----	-----	----	3.00	34.40	25.00	40.60
													51.03	24.88	24.08
		0	64.92	18.57	16.49	1.21		4.11							

* NUMBER = sample number DEPTH 1 = feet DEPTH 2 = meters Q = quartz in QFR F = feldspar in QFR R = rock fragments in QFR NON-QFR = miscellaneous transported constituents P/P+K = percent plagioclase in total feldspar A = authigenic components as percent of total rock M = matrix in total rock P = porosity in total thin section %F = percent feldspar in total rock SIZE = average grain size in closest quarter Phi units SRF = sedimentary rock fragments in RF VRF = volcanic rock fragments in RF MRF = metamorphic rock fragments in RF

APPENDIX 2.

Electron microprobe data

Feldspars

NUMBER MISS	DEPTH R	DEPTH O	N	Q	F	O	P/P+K	$\bar{X}$ AN	<AN	2 >AN	30	SIZE
PP-439	1454	443	211	82.50	14.70	2.80	0.59	13.30	15		0	1.50
PP-129	2728	831	239	77.00	15.90	7.10	0.49	22.50	14		36	2.00
PP-130	2741	835	190	76.30	17.40	6.30	0.46	31.10	0		47	2.25
PP-93	3300	1006	363	84.00	11.00	5.00	0.52	20.20	6		22	1.25
PP-95	3385	1032	156	75.00	19.20	5.80	0.37	13.90	18		0	1.50
PP-101	4632	1412	391	90.80	7.40	1.80	0.60	22.90	18		41	2.25
PP-57	4662	1421	131	86.20	9.20	4.60	0.44	28.00	0		43	2.25
PP-138	4704	1434	261	76.60	17.20	6.10	0.60	37.40	8		35	2.25
PP-109	5194	1583	196	71.40	20.40	8.20	0.50	33.00	6		63	2.75
PP-66	5345	1629	121	61.10	25.60	13.20	0.75	19.80	19		19	4.00
PP-17	5854	1784	145	66.20	26.20	7.60	0.66	13.60	32		18	4.00
PP-23	5876	1791					0.69	19.90	9		18	3.25
PP-196	6283	1915	397	85.10	10.10	4.80	0.31	20.90	18		36	1.50
PP-38	6334	1931	145	86.90	7.60	5.50	0.55	18.30	9		17	2.00
PP-74	6681	2036	187	67.40	22.50	10.20	0.55	31.50	10		30	3.00
PP-357	7040	2146	168	72.00	23.20	4.80	0.67	26.70	12		46	3.50
PP-119	7323	2232	181	63.50	21.00	15.50	0.70	21.50	22		37	2.50
PP-81	7456	2273	114	65.80	31.60	2.60	0.50	32.10	8		54	2.25
PP-826	8268	2520					0.36	25.30	0		50	2.00
PP-45	8587	2617	225	76.00	21.80	2.20	0.58	29.90	5		52	3.75
PP-173	8883	2708	254	75.60	16.10	8.30	0.62	24.20	8		46	2.75
PP-771	8983	2738					0.59	28.80	7		53	3.25
PP-54	9167	2794	160	68.10	26.30	5.60	0.67	17.90	5		25	4.00
PP-172	9747	2971					0.67	19.20	12		46	3.75
PP-343	10042	3061	115	60.90	34.80	4.30	0.56	28.00	10		50	3.75
PP-308	10929	3331	397	86.70	11.80	1.50	0.43	11.70	31		6	3.00
PP-247	10942	3335	324	84.90	8.90	6.20	0.67	12.80	25		25	2.75
PP-311	11896	3626	272	72.40	17.30	10.30	0.50	17.60	32		27	3.25
PP-176	12140	3700	193	73.60	20.70	5.70	0.75	26.50	3		45	3.75
PP-180	12163	3707	106	63.20	25.50	11.30	0.67	20.10	19		38	4.00
PP-183	12855	3918	240	83.30	10.40	6.30	0.82	17.50	11		11	3.25
PP-186	12859	3919					0.74	15.80	18		22	3.25
PP-263	13104	3994	271	81.90	13.70	4.40	0.63	15.10	20		30	2.25
PP-222	14033	4277	283	84.10	14.80	1.10	0.78	20.20	33		33	3.75
PP-270	14355	4375	340	88.50	9.10	2.40	0.87	13.40	15		15	3.50
PP-159	14357	4376	118	66.10	27.10	6.80	0.79	20.00	12		32	3.25
PP-275	14484	4415	160	65.00	25.00	10.00	0.73	25.70	7		37	3.00
PP-497	14823	4518	213	74.60	19.70	5.60	0.71	18.90	22		26	3.50
PP-157	14933	4552					0.85	15.10	17		14	3.50
PP-282	15369	4684	296	76.00	10.50	3.40	0.83	10.10	37		4	2.75
PP-288	16556	5046	241	80.10	12.40	7.50	1.00	6.70	40		0	4.00
		0	8095	75.68	17.89	6.14						

Appendix 2 continued

Clays <1 um

NUMBER	DEPTH	DEPTH	SIO2	AL2O3	TIO2	MGO	FEO	CAO	K2O	NA2O
PP-90	3265	995	57.58	24.44	.71	3.65	8.06	1.23	3.11	1.14
PP-56	4530	1381	55.07	25.17	.71	4.04	8.04	1.66	3.93	1.16
PP-62	5019	1530	56.58	26.29	.6	3.3	7.27	.94	4.23	.67
PP-71	6654	2028	57.46	26.02	.62	3.23	6.58	1.82	3.66	.49
PP-141	7259	2213	52.34	25.07	.74	3.92	6.89	.8	3.6	1
PP-77	7425	2263	57.25	25.63	.75	3.11	7.17	1.05	3.83	.63
PP-255	8055	2455	56.68	26.84	.63	3.03	6.53	1.72	3.33	1
PP-390	10817	3297	56.15	26.77	.7	3.2	7.47	1.37	3.64	.59
PP-393	11406	3477	56.46	26.73	.68	3.13	7.29	1.5	3.44	.61
PP-162	11476	3498	58.48	24.2	.68	3.31	6.88	1.04	4.17	1.12
PP-217	11842	3609	56.72	26.05	.68	3.09	7.26	.91	4.3	.81
PP-760	11912	3631	58.31	23.46	.69	3.8	7.99	.88	3.73	.99
PP-218	12329	3758	57	26.17	.74	3.1	6.97	.78	4.47	.71
PP-755	12343	3762	56.71	24.15	.75	3.6	8.87	.85	3.82	1.1
PP-262	13052	3978	57.87	26.71	.64	2.95	5.3	1.37	4.54	.56
PP-575	13785	4202	56.51	25.45	.75	2.94	8.18	1.59	3.53	.86
PP-158	14069	4288	56.76	27.04	.73	2.88	5.95	1.22	4.68	.65
PP-572	14115	4302	55.88	26.15	.7	3.14	7.39	2.1	3.85	.6
PP-269	14298	4358	55.97	28.81	.69	2.56	5.49	1.22	4.67	.46
PP-223	14310	4362	55.9	26.75	.71	3.11	6.71	1.1	4.78	.86
PP-159	14357	4376	55.49	26.41	.74	3.03	7.75	1.2	4.42	.8
PP-157	14933	4552	56.02	27.66	.68	2.95	6.23	.8	4.86	.68
PP-280	15169	4624	56.77	28.32	.74	2.64	5.65	.55	4.52	.74
PP-285	15997	4876	55.93	28.48	.76	2.51	5.64	1.1	4.9	.57

REFERENCES CITED

American Association of Petroleum Geologists and United States Geological Survey, 1976, Geothermal gradient map of North America, 1:5,000,000.

Andel, T. H. van, 1960, Sources and dispersion of Holocene sediments, northern Gulf of Mexico: p. 34-55 in, F.P. Shepard, F. B. Phleger, T. H. van Andel, (eds.), Recent Sediments, Northwest Gulf of Mexico: American Association of Petroleum Geologists, Tulsa, Oklahoma, 349 p.

Beard, J. H. and J. L. Lamb, 1968, The lower limit of the Pliocene and Pleistocene in the Carribean and Gulf of Mexico: Gulf Coast Association of Geological Societies Transactions, v. 18, p. 174-186.

Bernard, H. A., and R. LeBlanc, 1965, Resume of the Quaternary geology of the northwestern Gulf of Mexico province, pp.137-185, in H. E. Wright and D. G. Frey (eds.), Quaternary of the United States.

Berner, R. A., 1980, Early diagenesis, A theoretical approach: Princeton University Press, New Jersey, 241 p.

Berner, R. A. and G. R. Holdren, Jr., 1979, Mechanism of feldspar weathering--II. Observations from feldspars from soils: Geochimica et Cosmochimica Acta, v. 43, p. 1173-1186.

Bjorlykke, K., 1979, Cementation of sandstones: discussion: Journal of Sedimentary Petrology, v. 49, p. 1358-1359.

Boles, J. R., 1982, Active albitization of plagioclase, Gulf Coast Tertiary: Amer. Jour. Sci., v. 282, p. 165-180.

Boles, J. R. and S. G. Franks, 1979, Clay diagenesis in Wilcox sandstones of southwest Texas: implications of smectite diageneis on sandstone cementation: Journal of Sedimentary Petrology,

v. 49, p. 55-70.

Braunstein, J., J. A. Hartman, B. L. Kane, and J. H. van Amringe, (eds.), 1973, Offshore Louisiana oil and gas fields: New Orleans and Lafayette Geological Societies, 124 p.

Bruce, C. H., 1984, Smectite dehydration--its relationship to structural development and hydrocarbon accumulation in Northern Gulf of Mexico basin: American Association of Petroleum Geologists Bulletin, v. 68, p. 673-683.

Buffler, R. T., F. J. Shaub, R. Huerta, A. B. K. Ibrahim, and J. S. Watkins, 1981, A model for the early evolution of the Gulf of Mexico basin: Oceanologica Acta, Proceeding of the International Geological Congress Geology of Continental Margins Symposium, p. 129-136.

Buffler, R. T., J. L. Worzel, and J. S. Watkins, 1978, Deformation and origin of the Sigsbee Scarp--Lower continental slope, northern Gulf of Mexico: Offshore Technology Conference.

Caughey, C. A., 1975, Pleistocene depositional trends host valuble Gulf oil reserves (Parts I and II): Oil and Gas Journal, v. 36 and 37, p. 91-94, 240-247.

Chenoweth, P. A. and M. H. McBride, 1984, Formation correlation of the Gulf coast and Atlantic seaboard: Pennwell Maps, Tulsa, Oklahoma, 2 sheets.

Clayton, R. N. and T. K. Mayeda, 1963, The use of bromine pentafluoride in the extraction of oxygen from oxides and silicates for isotopic analysis; Geochimica et Cosmochimica Acta, v. 27, 43-52.

Curtis, C. D., 1978, Possible links between sandstone diagenesis and depth-related geochemical reactions occurring in enclosing mudstones: Geological Society of London Journal, v. 135, p. 107-117.

Davies, D. K. and W. R. Moore, 1970, Dispersal of Mississippi sediment in the Gulf of Mexico: Journal of Sedimentary Petrology, v. 40, p. 339-353.

Deuser, 1970, Extreme C/ C variations in Quaternary dolomites from the continental shelf: Earth and Planetary Science Letters, v. 8, p. 118-124.

Dickinson, W. R. and P. J. Coney, 1980, Plate tectonic constraints on the origin of the Gulf of Mexico: pp. 27-36 in R. H. Pliger (ed.), The origin of the Gulf of Mexico and the early opening of the central north Atlantic: Proceedings of a Symposium at Louisiana State University.

Dickinson, W. R. and C. A. Suczek, 1979, Plate tectonics and sandstone compositions: American Association of Petroleum Geologists Bulletin, v. 63, 2164-2182.

Elders, W. A., J. R. Hoagland, S. D. McDowell, and J. M. Hoagland, 1979, Hydrothermal mineral zones in the geothermal reservoir of Cerro Prieto: Geothermics, v. 8, p. 201-209.

Eslinger, E,, 1971, Mineralogy and oxygen isotope ratios of hydrothermal and low-grade metamorphic argillaceous rocks: unpub. Ph.D. dissertation, Case Western Reserve University, 205 p.

Eslinger, E. and S. M. Savin, 1973, Oxygen isotope geothermometry of the burial metamorphic rocks of the Precambrian Belt Supergroup, Glacier National Park, Montana: Geological Society of America Bulletin, v. 84, p. 2549-2560.

Fisher, R. S., 1982, Diagenetic history of Eocene Wilcox sandstones and associated formation waters, south-central Texas: unpub. Ph.D. dissertation, University of Texas at Austin, 185 p.

Folk, R. L., 1975, Glacial deposits identified by chattermark trails in detrital garnets: Geology, v. 3, p. 473-475.

Folk, R. L., 1980, Petrology of Sedimentary Rocks: Hemphill Publishing Co., Austin, Texas, 182 p.

Freed, R. L., 1981, Shale mineralogy and burial diagenesis of Frio and Vicksburg formations in two geopressured wells, McAllen Ranch area, Hidalgo County, Texas: Gulf Coast Association of Geological Societies Transactions, v. 31, 289-294.

Freed, R. L., 1982, Clay mineralogy and depositional history of the Frio Formation in two geopressured wells, Brazoria County, Texas: Gulf Coast Association of Geological Societies Transactions, v. 32, p. 459-463.

Fritz, P., P. L. Binda, F. E. Folinsbee, and H. R. Krouse, 1971, Isotopic composition of diagenetic siderite from Cretaceous sediments in Western Canada: Journal of Sedimentary Petrology, v. 41, p. 282-288.

Galloway, W. E., 1984, Hydrogeologic regimes of sandstone diagenesis: pp. 3-13 __in D. A. Macdonald and R. C. Surdam, (eds.), Clastic Diagenesis: American Association of Petroleum Geologists Memoir 37, 434 p.

Galloway, W. E., D. K. Hobday, and K. Magara, 1982, Frio Formation of the Texas Gulf coastal plain--depositional systems, structural framework, and hydrocarbon distribution: American Association of Petroleum Geologists Bulletin, v. 66. p. 649-688.

Gold, P. B, 1984, Diagenesis of Middle and Upper Miocene sandstones, Louisiana Gulf coast: unpub. M. A. thesis, University of Texas at Austin, 159 p.

Grim, R. E. and W. D. Johns, 1958, Clay mineral investigations of sediments in the mnorthern Gulf of Mexico: Clays and Clay Minerals, v. 327, 81-103.

Hangari, K. M., S. N. Ahmad, and E. C. Perry, Jr., 1980, Carbon and oxygen isotope ratios in diagenetic siderite and magnetite from upper Devonian ironstone, Wadi Shatti district, Libya: Economic Geology, v. 75, p. 538-545.

Holland, D. S., W. E. Nunan, D. R. Lammlein, and R. L. Woodhams, 1980, Eugene Island Block 330 Field,

offshore Louisiana, in Giant Oil and Gas Fields of the Decade, 1968-1978, American Association of Petroleum Geologists Memoir 30, Tulsa, p. 253-280.

Houghton, H. F., 1980, Refined techniques for staining plagioclase and alkali feldspars in thin section: Journal of Sedimentary Petrology, v. 50, p. 629-631.

Hower, J., E. V. Eslinger, M. E. Hower, and E. A. Perry, 1976, Mechanism of burial metamorphism of argillaceous sediment: I. Mineralogical and chemical evidence: Geological Society of America Bulletin, v. 87, p. 725-737.

Ibrahim, A. K., J. Carye, G. Latham, and R. T. Buffler, 1981, Crustal structure in Gulf of Mexico from OBS refraction and multichannel reflection data: American Association of Petroleum Geologists Bulletin, v. 65, p. 1207-1229.

Johns, W. D. and R. E. Grim, 1958, Clay mineral compositions of Recent sediments from the Mississippi River delta: Journal of Sedimentary Petrology, v. 28, p. 186-199.

Johnson, L. C., 1980, Structure and stratigraphy of an evolving salt ridge and basin complex, Louisiana contiental shelf: unpub. M. A. thesis, University of Texas at Austin, 120 p.

Jones, P. H., 1969, Hydrodynamics of geopressure in the northern Gulf of Mexico basin: Journal of Petroleum Technology, v. 21, p. 803-810.

Jones, P. H., 1975, Geothermal and hydrothermal regimes, Northern Gulf of Mexico basin: pp. 15-89, in M. H. Dorfman and R. W. Deller (eds.), First Geopressured Geothermal Energy Conference Proceedings, University of Texas at Austin, p.

Kehle, R. O., 1971, Geothermal survey of North America, 1971 annual progress report: Research Committee, American Association of Petroleum Geologists, unpublished report, 31 p.

Kharaka, Y. K., E. Callender, and R. H. Wallace, 1978, Geochemistry of geopressured-geothermal waters

from the Frio Clay in the Gulf Coast region of Texas: Geology, v. 5, p. 241-244.

Kinney, D. M., 1966, Geologic map of the United States, 1:7,000,000: U. S. Geological Survey.

Ladd, J. W., R. T. Buffler, J. S. Watkins, J. L. Worzel, and A. Carranza, 1976, Deep seismic reflection results from the Gulf of Mexico: Geology, v.4, p. 365-368.

Land, L. S., 1983, The application of stable isotopes to studies of the origin of dolomite and to problems of diagenesis in clastic sediments: in M. A. Arthur and others, Stable isotopes in sedimentary geology, Society of Economic Paleontologists and Mineralogists, Short Course No. 10.

Land, L. S., 1984a, Evidence for vertical movement of diagentic fluids, Texas Gulf coast: American Association of Petroleum Geologists Annual Meeting, Abstracts.

Land, L. S., 1984b, Frio sandstone diagenesis, Texas Gulf coast: A regional isotopic study: pp. 47-62.in.D. A. MacDonald and R. C. Surdam (eds.) Clastic diagenesis: American Association of Petroleum Geologists Memoir 37, 434 p.

Land, L. S. and S. P. Dutton, 1978, Cementation of a Pennsylvanian deltaic sandstone: isotopic data: Journal of Sedimentary Petrology, v. 48, p. 1167-1176.

Land, L. S. and K. L. Milliken, 1981, Feldspar diagenesis in the Frio Formation, Brazoria County, Texas Gulf Coast: Geology, v. 9, p. 314-318.

Land, L. S. and D. R. Prezbindowski, 1981, The origin and evolution of saline formation water, Lower Cretaceous carbonates, south-central Texas, U. S. A.: Journal of Hydrology, v. 54, p. 51-74.

Lindquist, S. J., 1977, Secondary porosity development and subsequent reduction, overpressured Frio Formation sandstone (Oligocene), south Texas: Gulf Coast Association of Geological Societies

Transactions, v. 27, p. 99-107.

Loucks, R. G., 1976, Pearsall Formation, Lower Cretaceous, south Texas--Depositional facies and carbonated diagenesis and their relationship to porosity: unpub. Ph.D. dissertation, University of Texas at Austin, 362 p.

Loucks, R. G., M. M. Dodge, and W. E. Galloway, 1984, Regional controls on diagenesis and reservoir quality in lower Tertiary sandstones along the Texas Gulf coast: pp. 15-45.in.D. A. MacDonald and R. C. Surdam, (eds.), Clastic Diagenesis: American Association of Petroleum Geologists Memoir 37, 434 p.

Lundegard, P. D., L. S. Land, and W. E. Galloway, 1984, Problem of secondary porosity, Frio Formation (Oligocene), Texas Gulf coast: Geology, v. 12, p. 399-402.

Lundegard, P. D., 1985, Carbon dioxide and organic acids: origin and role in diagenesis, Texas Gulf coast Tertiary: Ph.D. dissertation, University of Texas at Austin.

McCrea, J. M., 1950, On the isotopic chemistry of carbonates and a paleotemperature scale: Journal of Chemistry and Physics, v. 18, p. 849-857.

Martin, R. G., 1978, Northern and eastern Gulf of Mexico continental margin: stratigraphic and structural framework: pp. 21-42, in A. H. Bouma, G. T. Moore, and J. M. Coleman (eds.) Framework, facies, and oil-trapping characteristics of the upper continental margin, American Association of Petroleum Geologists Studies in Geology No. 7, 326 p.

Maynard, J. B., 1984, Composition of plagioclase feldspar in modern deep-sea sands: Relationship to tectonic setting: Sedimentology, v. 31, p. 493-501.

Merritt, L. C., 1980, Sandstone diagenesis of Olmos, San Miguel, and Upson formations (Upper Cretaceous, Northern Rio Escondido Basin, Coahuila, Mexico: unpub. M. A. thesis, University of Texas at Austin, 136 p.

Milliken, K. L., 1979, The silicified evaporite syndrome--two aspects of silicification history of former evaporite nodules from southern Kentucky and northern Tennessee: Journal of Sedimentary Petrology, v. 49, p. 245-256.

Milliken, K. L. and L. S. Land, 1982, Fluid dynamics for cap-rock formation in Gulf Coast: discussion: American Association of Petroleum Geologists Bulletin, v. 66, p. 2685-2687.

Milliken, K. L., L. S. Land, and R. G. Loucks, 1981, History of burial diagenesis determined from isotopic geochemistry, Frio Formation, Brazoria County, Texas: American Association of Petroleum Geologists Bulletin, v. 65, p.1397-1413.

Moore, C. H. and Y. Druckman, 1981, Burial diagenesis and porosity evolution, Upper Jurassic Smackover, Arkansas and Louisiana: American Association of Petroleum Geologists Bulletin, v. 65, p. 597-628.

Morton, R. A., C. M. Garrett, Jr., J. S. Posey, J. H. Han, and L. A. Jirik, 1981, Salinity variations and chemical compositions of waters in the Frio Formation, Texas Gulf Coast: Bureau of Economic Geology, Contract report to Dept. of Energy, No. DE-AC08-79ET17111.

Murray, G. E., 1960, Geologic framework of Gulf Coastal province of United States: pp. 5-33, in F. P. Shepard, F. B. Phleger, and T. H. VanAndel (eds.) Recent sediments, Northwest Gulf of Mexico, American Association of Petroleum Geologists, Tulsa, Oklahoma, 394 p.

Norwood, E. M. and D. S. Holland, 1974, Lithofacies mapping: A descriptive tool for ancient delta systems of the Louisiana outer continental shelf: Gulf Coast Association of Geological Societies Transactions, v. 24, p. 175-188.

O'Neil, J. R., R. N. Clayton, and T. K. Mayeda, 1969, Oxygen isotopic fractionation in divalent metal carbonates: Journal of Chemistry and Physics, v. 51, p. 5547-5558.

Parker, R. H., 1960, Ecology and distribution patterns of marine macro-invertebrates, northern Gulf of Mexico: pp. 302-337, in F.P. Shepard, F.B. Phleger, and T. H. van Andel, (eds.), Recent Sediments, Northwest Gulf of Mexico, American Association of Petroleum Geologists, Tulsa, Oklahoma, 394 p.

Pinsak, A. P. and H. H. Murray, 1960, Regional clay mineral patterns in the Gulf of Mexico: Proc. of Seventh National Conference on Clays and Clay Minerals, v. 7, p. 162-177.

Potter, P. E., D. Heling, N. F. Shimp, and W. A. Van Wie, 1975, Clay mineralogy of modern alluvial mud of the Mississippi River basin: Bull. Centre Rech. Pau SNPA, v. 19, p. 353-389.

Prezbindowski, D. R., 1981, Burial diagenesis, Edwards Formation, Lower Cretaceous, south-central Texas: unpub. Ph.D. dissertation, University of Texas at Austin, 235 p.

Rettke, R. C., 1976, Clay mineralogy and clay mineral distribution patterns in Dakota Group sediments, northern Denver basin, eastern Colorado and Western Nebraska: unpublished Ph.D. dissertation, Case Western Reserve University, 135 p.

Reynolds, R. C. and J. Hower, 1970, The nature of interlayering in mixed-layer illite-montmorillonites: Clays and Clay Minerals, v. 18, p. 25-36.

Roberts, H. H. and T. Whelan III., 1975, Methane-derived carbonate cements in barrier and beach sands of a subtropical delta complex: Geochimica et Cosmochimica Acta, v. 39, p. 1085-1089.

Russell, K. L. and others, 1966, Marine dolomite of unusual isotopic composition: Science, v. 155, p. 189-191.

Scafe, D. W. and G. W. Kunzel, 1971, A clay mineral investigation of six cores from the Gulf of Mexico: Marine Geology, v. 10, p. 69-85.

Schlager, W. and others, (in press), D. S. D. P. Leg 77--Early history of the Gulf of Mexico.

Shapiro, L., 1975, Rapid analysis of silicate, carbonate, and phosphate rocks--revised edition: U. S. Geological Survey Bulletin 1401, 76 p.

Sharp, J. M., Jr. and P. A. Domenico, 1976, Energy transport in thick sequences of compacting sediment: Geological Society of America Bulletin, v. 87, p. 390-400.

Shepard, F. P., 1960, Rise of sea level along northwest Gulf of Mexico: pp. 338-344, in F. P. Shepard, F.B. Phleger, and T. H.Van Andel (eds.), Recent Sediments, Northwest Gulf of Mexico, American Association of Petroleum Geologists, Tulsa, Oklahoma, 394 p.

Slator, D. S., 1980, Sandstone diagenesis and its variation with deltaic depositional environments, upper Cretaceous, southern Rio Escondido Basin, Coahuila, Mexico: unpub. M. A. thesis, University of Texas at Austin, 195 p.

Spillers, J. P., (ed.), 1962, Salt domes of south Louisiana, Volume II: New Orleans Geological Society, New Orleans, Louisiana, 107 p.

Spindler, W. M., 1977, Structure snd stratigraphy of a small Plio-Pleistocene depocenter, Louisiana continental shelf: Gulf Coast Association of Geological Societies Transactions, v. 27, p. 180-196.

Stude, G. R., 1978, Depositional environments of the Gulf of Mexico South Timbalier Block 54, salt dome and salt dome growth models: Gulf Coast Association of Geological Societies Transactions, v. 28, p. 627-646.

Syers and others, 1968, Quartz isolation from rocks, sediments, and soils for determination of oxygen composition: Geochimica et Cosmochimica Acta, v. 32, p. 1022-1025.

Trevena, A. S. and W. P. Nash, 1981, An electron microprobe study of detrital feldspar Journal of

Sedimentary Petrology, v. 51, p. 137-150.

Waguespack, S. J., (ed.), 1983, Salt domes of south Louisiana, Vol. III: New Orleans Geological Society, New Orleans, Louisiana, 142 p.

Weiss, B. N., 1979, Wave-dominated deltaic systems of the upper Cretaceous San Miguel Formation, Maverick Basin, south Texas: unpub. M. A. thesis, University of Texas at Austin, 116 p.

Wenner, D. L. and H. P. Taylor, 1971, Temperature of serpentinization of ultramafic rocks based on 180/160 fractination between coexisting serpentine and magnetite: Contributions to Mineralogy and Petrology, v. 32, p. 165-185.

Winker, C. D., 1981, Cenozoic shelf margins, northwestern Gulf of Mexico basin: pp. 74-82, in B. F. Perkins, W. P. S. Ventress, and M. B. Edwards (eds.) Recognition of shallow-water versus deep-water sedimentary facies in growth structure-affected formations of the Gulf Coast Basin, 2nd., Annual Research Conference, Gulf Coast Society of Economic Paleontologists and Mineralogists, 85 p.

Woodbury, H. O., I. B. Murray, P. J. Pickard, and W. H. Akers, 1973, Pliocene and Pleistocene depocenters, outer continental shelf, Louisiana and Texas: American Association of Petroleum Geologists Bulletin, v. 57, p. 2428-2439.

Woronick, R. E. and L. S. Land, (in press), Late burial diagenesis, lower Cretaceous Pearsall and Lower Glen Rose formations, south Texas: Society of Economic Paleontologists and Mineralogists Special Publication No. 36.

Worzel, J. L. and C. A. Burk, 1978, The margins of the Gulf of Mexico: pp. 403-419 in Geological and Geophysical Investigations of Continental Margins, American Association of Petroleum Geologists Memoir 29.

Yeh, H.-W. and S. M. Savin, 1977, Mechanism of burial metamorphism of argillaceous sediments: 3. Oxygen isotopic evidence: Geological Society of America Bulletin, v. 88, p. 1321-1330.

The vita has been removed from the digitized version of this document.